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UNUSUAL CURRENTS IN GLACIAL LAKE MISSOULA, MONTANA

BY J. T. PARDEE

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ABSTRACT

The area submerged by Glacial Lake Missoula includes several intermontane basins and constricted interconnecting valleys or "narrows" that drain to a single outlet, the Clark Fork River. A sudden failure of the ice dam that blocked this valley, near the Idaho-Montana State line, caused unusually large and rapid currents through the narrows and wind gaps in the partly submerged rim of Camas Prairie basin. Evidences of such currents include commensurate, but otherwise ordinary, effects of streams confined to rocky channels and the unique giant ripple marks. At its high stage the lake is roughly estimated to have held more than 500 cubic miles of water of which nearly three-fourths was stored above a constricted part of the Clark Fork Valley called the Eddy Narrows. Calculations based on available incomplete data indicate a flow through the Eddy Narrows that reached a maximum of 9.46 cubic miles per hour. Whether the lake was completely drained at that time has not been determined, but a later set of beaches testifies that the basin held a lake soon after the rapid outflow. Apparently the final draining was gradual.

INTRODUCTION

In the Pleistocene much of the Clark Fork drainage basin in northwestern Montana (Fig. 1) was briefly submerged by "Glacial Lake Missoula" (Pardee, 1910). A lobe of the Cordilleran glacier advanced southward along the Purcell Trench into northern Idaho where it occupied the basin of Pend Oreille Lake and crowded up the valley of the Clark Fork River to a position near the Montana State line. The water thus impounded in the Clark Fork Valley rose to a maximum altitude of 4150 feet or about 2000 feet higher than the floor of the valley just above the dam. On the east, south, and west it was held in by mountains, and on the north possible outflow channels were blocked by other lobes of Cordilleran ice.

The ice dam is thought to have failed, permitting a sudden large outflow. The consequent depression of the water surface in the arm

of the lake above the dam caused huge and rapid currents in the narrow valleys and passes connecting this arm with the wider depressions in the lake basin farther east. The records of these currents are described. The interpretation of certain gravel deposits and denuded

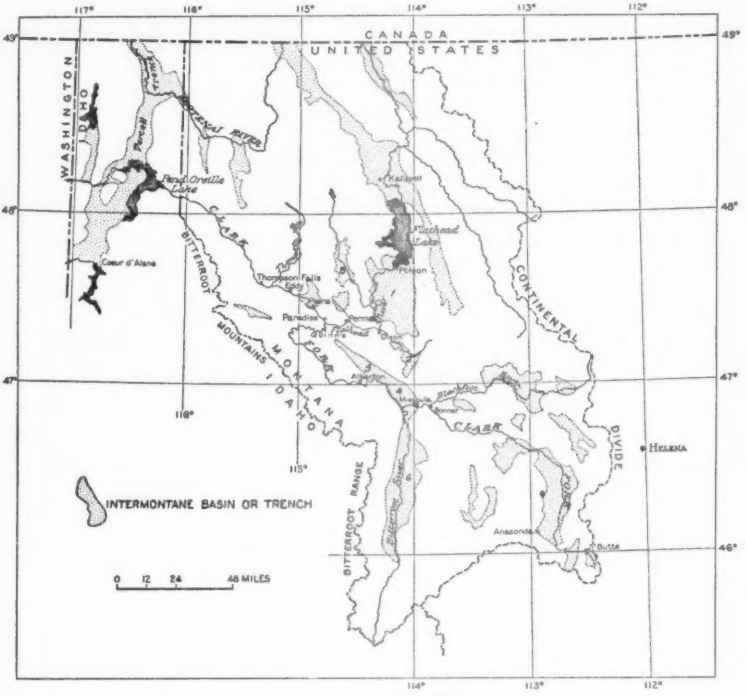


FIGURE 1.—*Index Map*

1. Mission Valley.
2. Jocko Valley.
3. Little Bitterroot Valley.
4. Missoula Valley.
5. Ninemile Valley.
6. Bitterroot Valley.

surfaces in the Perma and Paradise narrows as the effects of such currents differs from that of Davis (1921) who regarded them, respectively, as "moraines" and "ice scoured slopes." Field work by W. C. Alden and the writer has failed to uncover any evidence of glacial origin.

Not long after the rapid outflowing ceased the basin again held a lake. So far it is not clear whether this body was part of the original ponded water whose escape had been halted or a new filling. In contrast to the earlier outrush, final draining seems to have been very slow.

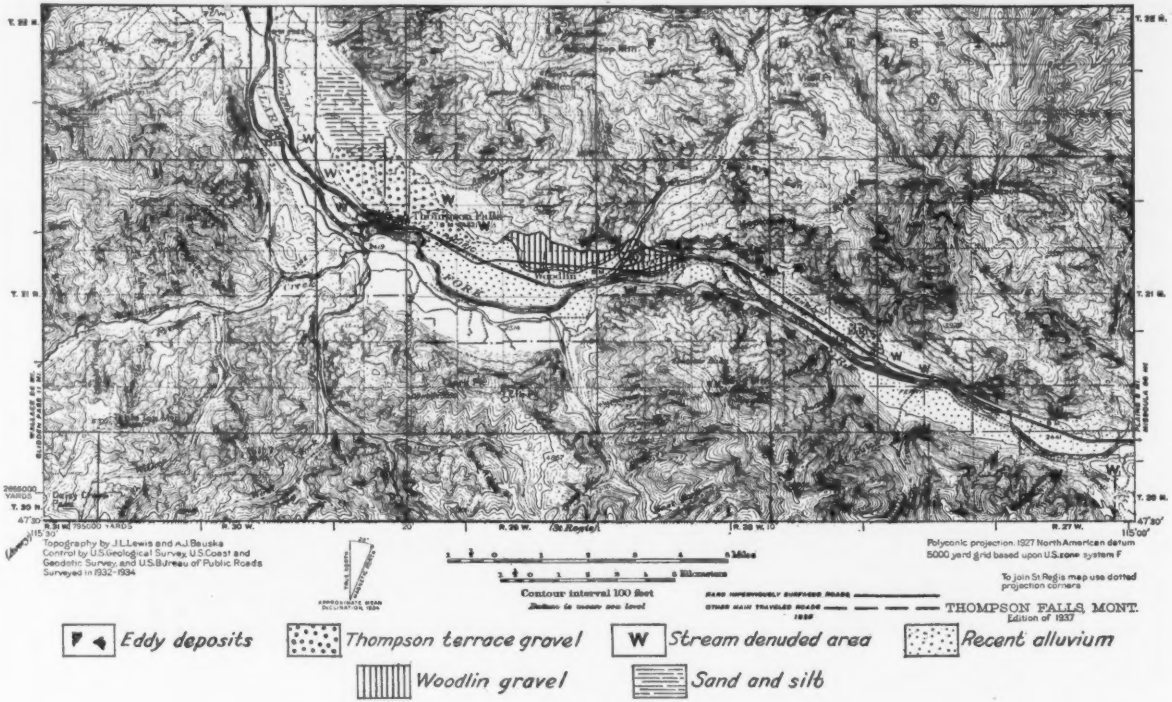

Topography by J.L.Lewis and A.J.Bauske
Control by U.S.Geological Survey, U.S.Coast and Geodetic Survey, and U.S.Bureau of Public Roads
Surveyed in 1932-1934
Contour interval 100 feet
Polyconic projection. 1927 North American datum
5000 yard grid based upon U.S.zone system F
To join St Regis map use dotted projection corners
THOMPSON FALLS, MONT.
Edition of 1937
Eddy deposits
Thompson terrace gravel
Stream denuded area
Recent alluvium
Woodlin gravel
Sand and silt

FIGURE 2.—*Eddy Narrows and Thompson Falls Valley*

Reproduced from Thompson Falls, Montana, topographic sheet of the U. S. Geol. Survey. (The term "eddy deposits" in the legend is equivalent to "high eddy deposits" in the text.)

ACKNOWLEDGMENTS

The field work for this paper was facilitated by the use of aerial photographs of the United States Forest Service without which much of the evidence presented could not have been obtained. To W. C. Alden of the United States Geological Survey the writer is indebted for collaboration in the field work and to W. W. Rubey and W. B. Langbein of the same bureau for calculations of stream flow.

TOPOGRAPHY

BASINS

Although the region of Glacial Lake Missoula is occupied chiefly by mountains, the topography is diversified by a number of wide valleys (Fig. 1) or intermontane basins.

The basins are enclosed by generally rough and steep mountains standing from 2000 to 4000 feet higher. Exceptionally high rugged ranges west of the Bitterroot Valley (Bitterroot Range) and east of Mission Valley (Mission Range) rise abruptly to serrate crests 5000 to 6000 feet above the adjacent lowlands. Some of the basins are isolated except for narrow outlet valleys that join them in a common drainage system. Others are more closely connected in irregular groups and in certain linear series known as trenches. Most of the individual basins are one-fifth to half as wide as they are long. Their floor areas range from about 50 square miles in Camas Prairie to 400 square miles or more in Mission Valley and consist mainly of gently sloping plains that merge with a medial flat or flood plain or end as terraces above it. Several of these basins were submerged by the central and eastern parts of the glacial lake. Among them the Mission, Jocko, and Little Bitterroot valleys (Figs. 1, 3) form a closely spaced group drained by the Flathead River. Another group that drains into the Clark Fork River above the mouth of the Flathead includes the Missoula, Ninemile, and Bitterroot valleys. Two other relatively small basins submerged in the central part of the lake are of particular interest in its history: Plains Valley (Fig. 9; Pl. 1, fig. 1) along the Clark Fork River a few miles below the mouth of the Flathead, and Camas Prairie, east of Plains Valley and a short distance north of Perma.

NARROWS

Clark Fork River and its tributaries drain the basins through comparatively narrow valleys which in certain places become severely constricted and gorgelike narrows. The 70-mile stretch of the Clark Fork Valley between the Missoula-Ninemile-Bitterroot group of basins and the Plains Basin generally is not more than 1 or 1½ miles wide

and is enclosed by steep, rugged mountains 2000 to 4000 feet high. Parts narrowed to half a mile or less occur near Alberton (Alberton Narrows), above Quinn's Springs (Quinn's Narrows), and below the

FIGURE 3.—*Glacial Lake Missoula*
High Stage

mouth of the Flathead River from Paradise to Plains (Paradise Narrows). The Clark Fork Valley throughout 10 miles below Plains Basin is again less than a mile wide (Eddy Narrows) between steep, rugged slopes 4000 feet high (Fig. 2). The valley widens abruptly at the lower end of this stretch and from there for about 70 miles to the northwest to the Idaho State line is generally straight and averages 2 and 4 miles in width. The Cabinet Mountains northeast of this

FIGURE 1. PLAINS VALLEY
One of smaller intermontane basins. Looking west into Eddy Narrows, the outlet. Timbered terrace in background is part of coarse gravel deposit. Cultivated area of Recent alluvium.

FIGURE 2. WILLS CREEK PASS
Looking north (upstream). Ponded channel depression in foreground. Projecting rock ledges with appended gravel deposits.

FIGURE 3. SOUTH SIDE OF EDDY NARROWS
Near Outlaw Gulch. Severely denuded spur at left. Normal waste mantle begins above 1000 feet.

FIGURE 1. LOOKING WESTWARD (downstream)
About 1 mile above Perma. Denuded rock bench at left separated from cliff by trench partly filled with Recent talus.

FIGURE 2. ENTRANCE TO NARROWS
Stream-denuded surfaces and channel excavations with ponds (B) and gravel deposits. From aerial photos by U. S. Forest Service.

FIGURE 3. DENUDED ROCK BENCH AT PERMA.

PERMA NARROWS AND ROCK BENCH

part of the valley are steep and regular and are 3000 to 5000 feet higher than the valley. The Bitterroot Mountains opposite are lower and less regular. Beyond the State line the Clark Fork Valley opens into the Purcell Trench which in turn joins the Spokane Valley.

The outlet of the Mission group of basins, the valley of the Flathead River, narrows from 2 miles near Dixon to less than a mile at McDonald Station 7 miles downstream (McDonald Narrows). At Perma, 6 miles farther down, the valley becomes still narrower. From Perma to the mouth of the Flathead River, where it merges with the Paradise Narrows, it averages less than half a mile in width (Perma Narrows). As the width of the valley decreases, the enclosing slopes become higher and more rugged. Below Perma they rise 3000 feet or more with slopes characterized by cliffs and bare rock ledges (Pl. 2, fig. 1).

The river bed generally occupies half or more of the valley floor within the different narrows. The remainder of the floor is formed chiefly of low alluvial terraces and an occasional projecting rock knob or ridge. In the adjoining wider stretches of the Clark Fork Valley the terraces are larger and commonly stand 100 to 200 feet above the river. Other pertinent topographic features described in the section on evidences of the unusual currents are wind gaps in the rim of Camas Prairie Basin.

GEOLOGY

BEDROCK FORMATIONS

The region of Glacial Lake Missoula is largely underlain by pre-Cambrian Belt sedimentary formations which, with some intrusive granitic bodies, compose the mountains and extend beneath the Tertiary and later sediments in the lowlands. In the Camas Prairie and Plains basins and the different narrows, the older rocks consist mainly of dark-gray argillites and sandstones of the Prichard formation and Ravalli group, and of a number of diorite dikes and sills. The rocks are deformed by large north to northwestward trending folds and faults and are closely jointed as a rule. Most of the finer-grained beds separate into thin slabs along cleavage planes generally parallel to the bedding. In addition the rocks are traversed by shear zones represented at the surface by notches and trenches, conspicuous on the sides of Perma and Paradise narrows.

These rocks weather to a rubble and finally to a fine-grained sand-clay soil. Above or beyond the area flooded by the lake rubble and soil mantle most of the level and moderate slopes to a depth of several feet, with only particularly resistant beds or dikes projecting through it. Close spacing of joint and cleavage planes produces a rather fine shingle talus with comparatively few large slabs or blocks.

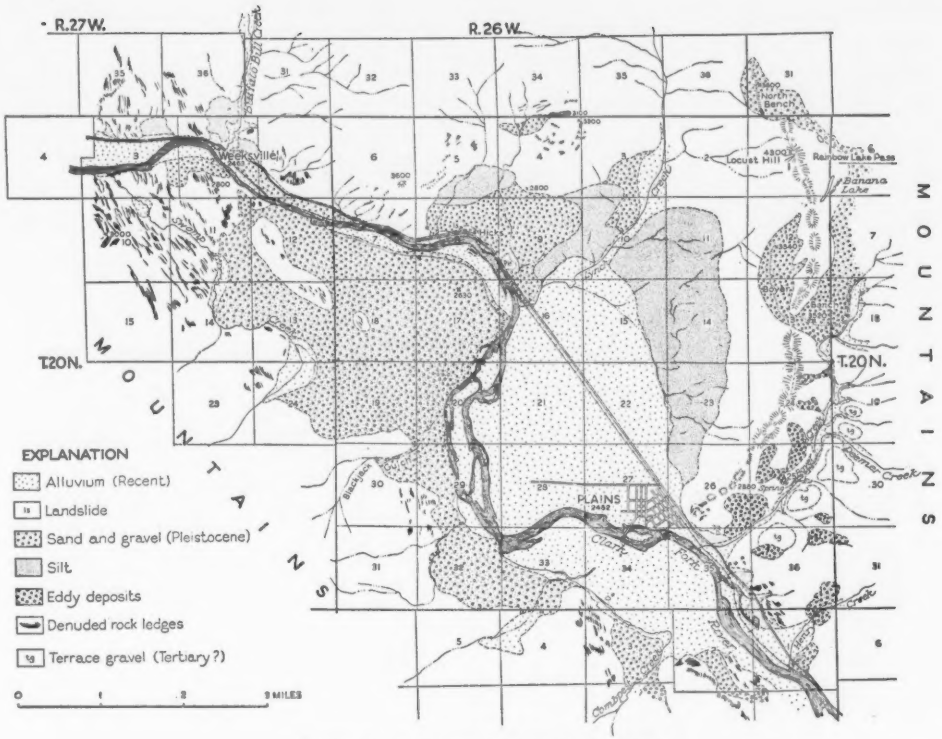


FIGURE 4.—*Southern Part of Plains Basin*

Denuded rock ledges indicated by lines parallel to their trends. The landslide indicated in the legend is on the bank of the river in Sec. 8, T. 20 N., R. 26 W.

On slopes that faced the wider reaches of water the mantle was more or less modified by wave action as described under the section on Beaches. In the narrows the mantle was largely removed by currents during the outflow.

TERTIARY (?) DEPOSITS

Tertiary sediments, extensive in the Missoula-Ninemile and Bitterroot basins, have not been positively identified within the submerged basins to the north and west. In those areas deposits unconformably on the bedrock which may be Tertiary include: (1) water-laid volcanic tuff at the north of Little Bitterroot Basin; (2) "rusty" gravel and sandstone exposed in road cuts and ditches along the foot of the Bison Range; (3) partly decomposed gravel exposed by a road cut in a bench east of Boyer Creek in Plains Basin; and (4) a small patch of gravel with smooth quartzite cobbles on the ridge west of Camas Creek about 2 miles north of Perma.

DEPOSITS ASSOCIATED WITH THE GLACIAL LAKE

Moraines and erratics.—The Pleistocene formations are comprehensively described by W. C. Alden (manuscript in preparation). Among those most closely related to the glacial lake are moraines of contemporaneous mountain glaciers and Cordilleran ice lobes. Several small moraines within the margin of the lake (Fig. 3) were deposited in the Mission and Jocko valleys by glaciers from the Mission Range and Missoula Hills on the east, and in the Bitterroot Valley by glaciers from the Bitterroot Range on the west. Larger moraines were formed at the fronts of the Cordilleran ice lobes that held back the lake on the north. The largest, the Polson moraine at the foot of Flathead Lake, marks the southern limit of the ice which held the arm of the glacial lake in the Mission Valley. Other moraines were deposited by Cordilleran ice lobes in Ninemile Prairie on the Blackfoot River, north of Big Prairie on the Thompson River, and in the valley of Bull River a few miles above its junction with the Clark Fork. Somewhat modified morainal material deposited by the lobe that blocked the Clark Fork Valley is exposed locally near the Idaho State line. Presumably it underlies and thus forms a peculiar sand-covered pitted terrace south of the river at Cabinet Gorge.

Erratics, evidently transported by icebergs from the different glaciers that entered or touched the lake, were distributed irregularly through the submerged areas.

A large body of fine-textured well-washed gravel (Fig. 2) apparently outwash from the retreating ice lobe in Thompson River Valley, forms 50-foot terraces in Clark Fork Valley above and below the entrance of Thompson River.

...parallel to their trends. The landslide indicated in the legend is on the bank of the river in Sec. 8, T. 20 N., R. 26 W.

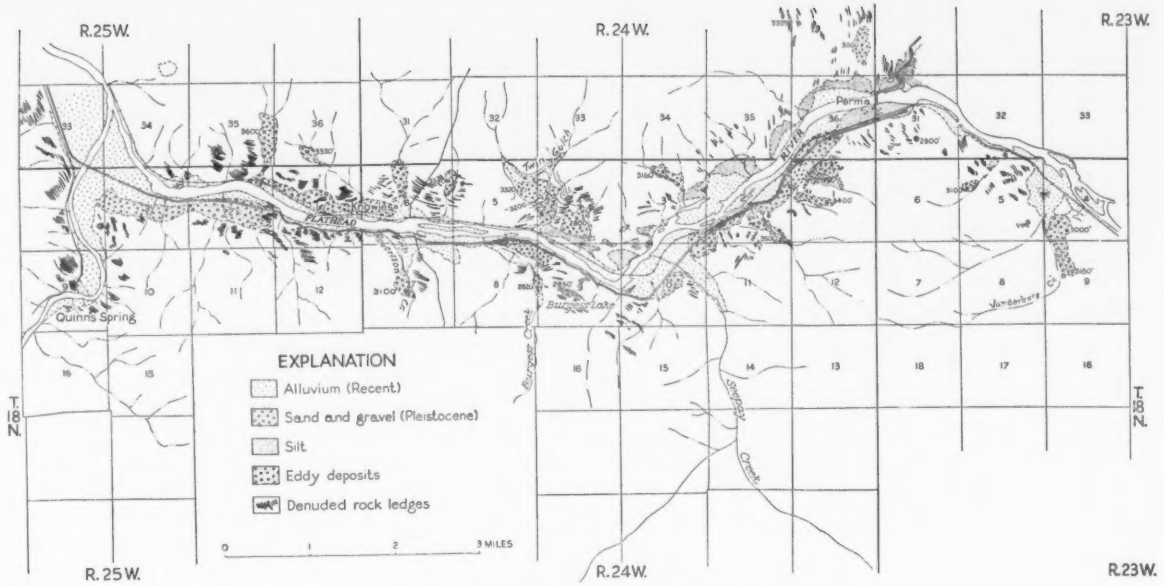


FIGURE 5.—*Perma Narrows*

Denuded rock ledges indicated by lines parallel to their trends.

Silt.—Light-colored varved silt, composed of glacial rock flour derived mostly from the lobes of Cordilleran ice, forms discontinuous terraces 50 to 100 feet high in most of the submerged basins. The silt is confined

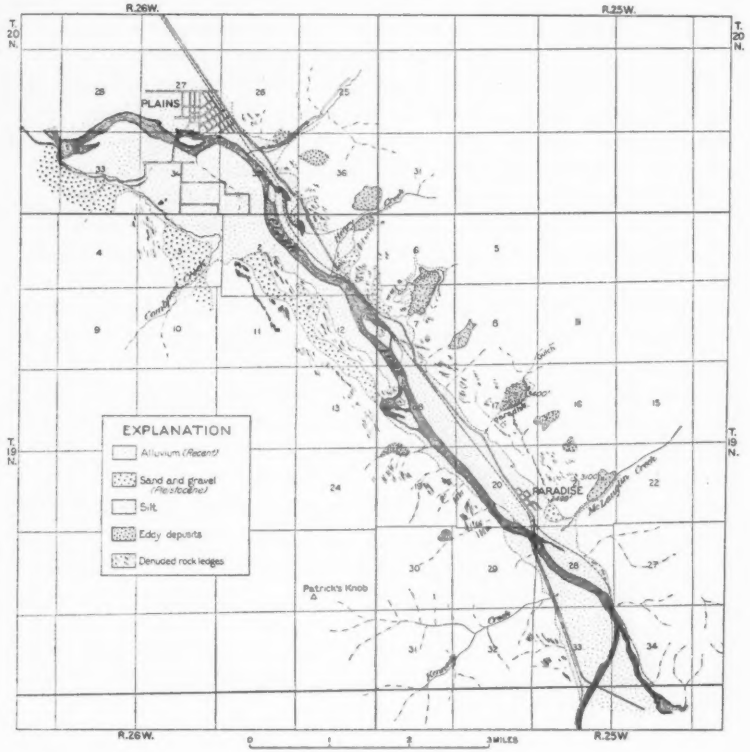


FIGURE 6.—*Paradise Narrows*
Denuded rock ledges indicated by lines parallel to their trends.

to the lower parts of the basins and thins out upslope at the height of the terraces which evidently mark its original upper limit. The Bitterroot Basin and, above the mouth of the Blackfoot River, the valley of the Clark Fork received meltwater from local mountain glaciers only, and in both areas little silt was deposited. In the Missoula-Ninemile Basin, however, silt derived apparently from large glaciers in the Blackfoot Valley forms extensive terraces 50 feet or more high between Missoula and Huson. In the Jocko Basin comparatively small discontinuous

terraces are composed of silt that came chiefly from glaciers washed by the arm of the lake in upper Jocko Valley. Silt discharged by the lobe of Cordilleran ice that filled the basin of Flathead Lake occupies large areas in the lower parts of the Mission and Little Bitterroot basins and forms discontinuous terraces up to 50 feet in height in the valley of the Flathead River (Pl. 7, fig. 3) to the Perma Narrows.

The silt in Little Bitterroot Basin is exposed by stream trenching to a depth of about 100 feet and is found to 200 feet in the middle of the basin in wells (Meinzer, 1916). In the Plains Basin, silt carried from either or both the Mission and the Missoula-Ninemile basins forms terraces about 100 feet high in a large area north of the river (Fig. 4).

In the Perma and Paradise narrows limited exposures of the silt occur (Figs. 5, 6) in protected places, and in the other narrows none has been observed. The somewhat wider stretches of the Clark Fork Valley between the Alberton Narrows and Quinn's Narrows and from the Eddy Narrows to the Idaho State line appear to contain silt in places, but its extent and relations have not been determined.

Deposition of the silt apparently was limited to the lower parts of the basins by conditions similar to those observed at Lake Mead above Boulder Dam on the Colorado River and in other reservoirs; owing to its higher specific gravity, slowly moving turbid water sinks and flows along the bottom without mixing with the overlying clarified water (Grover, *et al.*, 1938). In Lake Missoula the silt, in order to reach Plains Basin, must have been thus carried 50 to 75 miles from its sources in glaciers discharging their meltwater in the Mission and Missoula-Ninemile basins. No silt was deposited in Camas Prairie because none of the entrances to that basin were low enough to divert the layers of turbid water passing down the adjacent Little Bitterroot Basin and the Perma Narrows.

In Little Bitterroot Basin upper and lower silt formations separated by sand layers, and, at one place by gravel, are indicated by well records (Meinzer, 1916, p. 15). The partings occur at depths of 80 to 150 feet, and the underlying silt is 100 feet or more thick.

Deltas.—At all the different still stands of the lake, deltalike deposits of alluvium were doubtless formed at the upper limits of the arms extending into the many tributary valleys. The level of the water surface shifted rapidly so that individual deposits generally must have been rather small. Also, as they were exposed to later stream attack, few would remain sufficiently unmodified to be recognized. Gravel deposits in the Vermilion River Valley that may be the remnants of exceptionally large deltas begin at Cataract Creek about 5 miles from the mouth of

the valley and at Sims Creek about 3 miles farther up. The deposit at Cataract Creek forms a terrace 500 feet high along the south side of Vermilion River. From an altitude of 3200 feet at the lower end the surface ascends gradually upstream. The deposit consists of gravel overlying glacial till, which was observed at one place by W. C. Alden (personal communication). On the opposite side of Vermilion River and about a mile above Cataract Creek is a similar terrace in Lyons Gulch. A small undrained area at the back of this terrace indicates that the deposit was formed by currents turning out of the valley of Vermilion River. The terrace at Sims Creek is about 300 feet high, and the surface rises gradually upstream from an altitude of about 3550 feet at the terrace front.

A gravel filling in McLaughlin Gulch east of Paradise apparently was formed when the lake surface was at 3200 feet. About 1½ miles from the mouth of the gulch this deposit begins as a terrace about 30 feet high. Upstream it narrows and ends within half a mile because of the more steeply ascending bedrock channel of the gulch. The deposit is partly trenched by the stream and shows rather vague top-set bedding. Remnants of other gravel fillings apparently related to still stands of the lake extend up the valley of the St. Regis River from points above Taft at 3800 feet altitude and near Borax at 4100 feet. Both deposits rise gradually upstream from low terrace fronts.

Beaches.—At many places on the untimbered sides of the basins up to an altitude of 4150 feet the horizontal markings of closely spaced lake beaches are plainly visible from moderate distances. Within 1000 feet of the limiting altitude as many as 40 may be counted in favorable exposures (Pl. 5, fig. 1). Thus different still stands were so closely spaced that practically all of a given slope below 4150 feet was exposed to the waves which were most effective, of course, on the higher parts of a slope because they faced the wider stretches of water.

Scattered observations indicate that the beaches consist of a generally slight terracelike indentation thinly covered by cleanly washed and sorted gravel. On slopes that faced the wider arms of the lake the terrace or bench, as it rounds projecting spurs, may be several yards wide, and its related bars in adjacent re-entrants may be measured in cubic yards. Other conditions being equal, the beaches are not only more extensive around the larger basins but appear to be more deeply indented around their southern and eastern sides as if the prevailing winds were from the north and west. In many places rows of loose stones are ridged along the backs of the beaches where, presumably, they were shoved by expanding lake ice.

The beaches are less prominent in the smaller basins and become inconspicuous or absent in the valleys that held the narrower arms of the lake. In the relatively narrow valley of the Clark Fork River above Missoula they may be detected on hills west of Harvey Creek and west of Tyler Creek at altitudes between 3900 and 4150 feet. In the somewhat wider valley below Thompson Falls beaches may be observed between 2700 feet and 3500 feet altitude on partly bare slopes south of Tuscor and northeast of Noxon. Within the narrows beaches have not been identified.

RECENT ALLUVIUM

Within the basins recent alluvium forms flood plains and low terraces along the streams. In the narrows and the relatively narrow stretches of Clark Fork Valley, particularly below Thompson Falls, such features are absent or negligible because the river generally occupies a steep-sided inner trench.

EVIDENCES OF THE UNUSUAL CURRENTS

CHANNELED WIND GAPS IN RIM OF CAMAS PRAIRIE BASIN

General statement.—Camas Prairie Basin north of Perma is almost 12 miles long from north to south and 5 or 6 miles wide in the middle. Its floor descends gradually from an altitude of about 3100 feet at the north to about 2900 feet at the south and includes a flat from which the surface rises gradually to the moderately steep slopes of the surrounding mountains. At the high stage of the glacial lake (Fig. 3), the mountains on the east and west sides of the basins stood well above the water, but those on the north and south were almost entirely submerged. The arm of the lake in this basin was confluent, therefore, with the arms in Little Bitterroot Basin on the north and Perma Narrows on the south. In addition a strait in Rainbow Lake Pass connected it with the arm in Plains Basin on the west. When the water surface fell 500 feet or to about 3650 feet this pass became dry, and the northern rim partly emerged, thus dividing the water connection to the Little Bitterroot arm among four gaps. As the lake shrank further these gaps became dry, Duck Pond and Big Creek passes at about 3500 feet altitude, Markle Pass at 3400 feet, and finally Wills Creek Pass at 3300 feet. A wide gap at the south emerged when the lake fell about 300 feet more, and, except for the narrow outlet trench east of this gap, the water connection between the basin and Perma Narrows ceased.

Gaps in north rim.—Although they begin and end as it were "up in the air", the four gaps in the northern rim of Camas Prairie Basin carried large and vigorous streams. In Duck Pond Pass a marshy pond

with coarse gravel beyond it, and a bare rock hump, resemble a basin and succeeding bar formed by a south-flowing current plunging over the obstructing hump. Markle Pass and Wills Creek Pass have channel

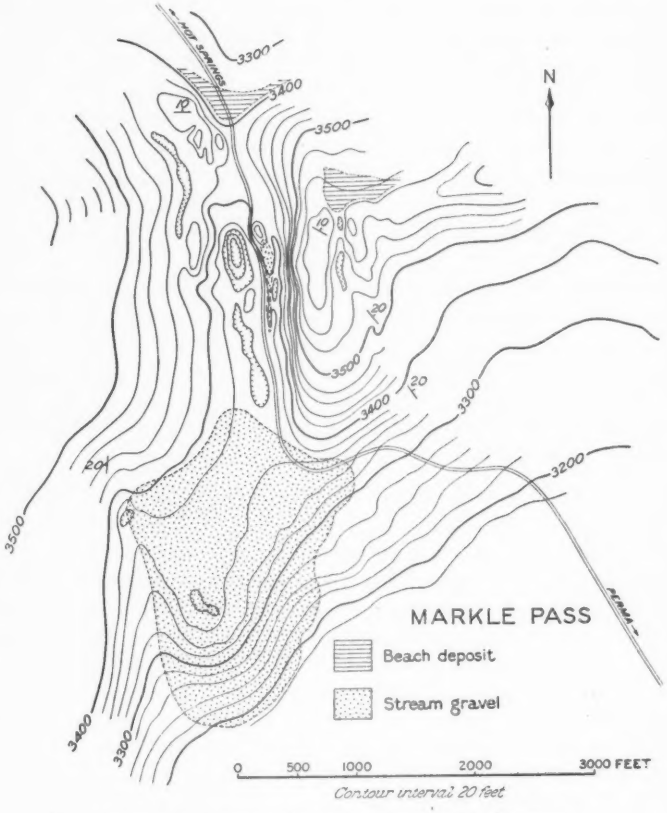


FIGURE 7.—*Topography and geology of Markle Pass*
Areas not patterned represent tilted Belt rocks.

excavations and discontinuous gravel deposits associated with projecting knobs and ridges. In Markle Pass (Fig. 7) an undrained depression 50 feet deep adjoins a prominent knob to which a gravel train is attached on the downstream side. East of the knob Recent talus partly fills another channel, and beyond that, to a height of 100 feet or more, the surface is denuded and channeled. South of the pass

coarse gravel with scattered large angular fragments plucked from ledges within has the form of a bar about 50 acres in extent built out to a height of 150 feet over the slope descending to the basin.

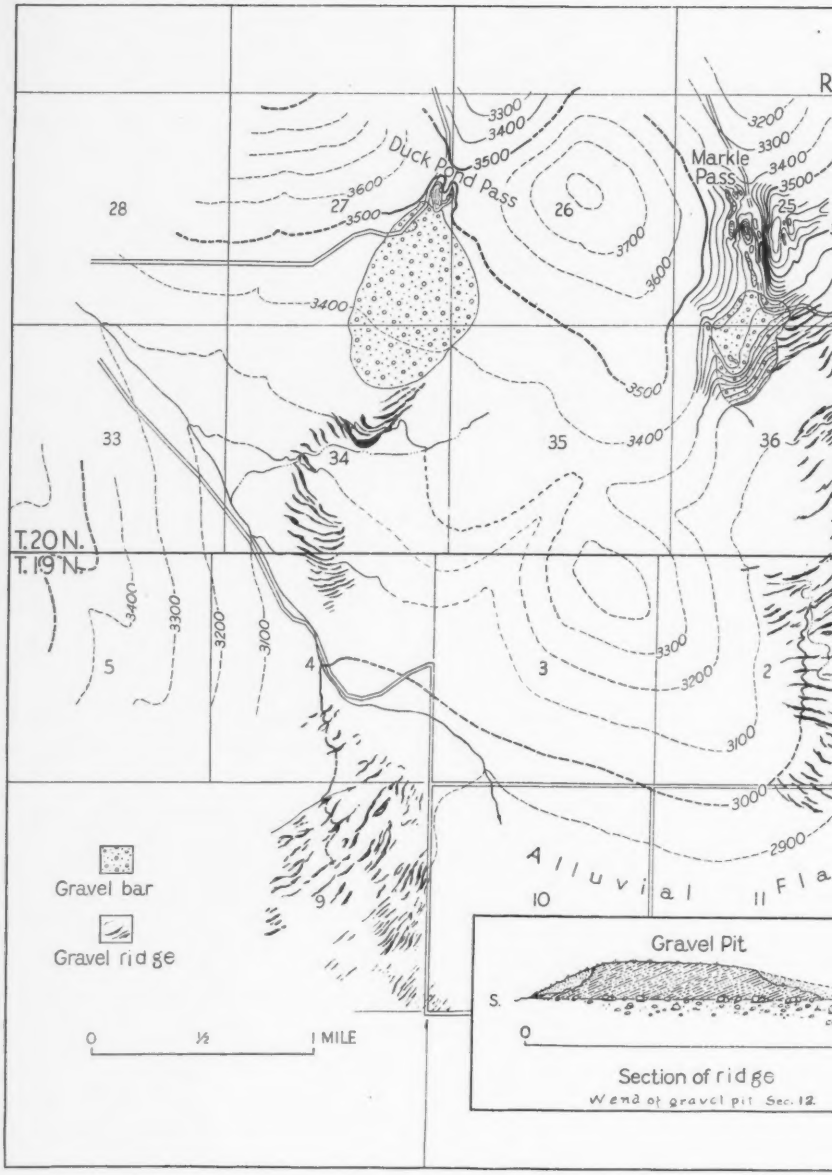
In Wills Creek Pass, channel features are developed on a still larger scale. Along the base of a steep slope at the east side is a pond 1000 feet long. On the west side (Pl. 1, fig. 2) are channels with undrained hollows and bare angular knobs and ridges projecting above discontinuous deposits of gravel. South of the pass a deltaic bar is built out for 1 mile on the slope descending to the basin (Pl. 3). The bar is 200 feet high at the southern end and like the deposit in front of Markle Pass is composed of coarse gravel with large angular fragments plucked from ledges within the channel. In front of Big Creek Pass a relatively small, almost circular barlike gravel deposit is built over a steep slope.

Apparently the inflowing currents were caused by a lowering of the water surface in the arm held by Camas Prairie Basin relative to the surface in Little Bitterroot Basin. Since the channels show no after effects of dwindling currents the flows must have ceased rather abruptly. The evidences described indicate actively eroding currents at least 30 or 40 feet deep in Duck Pond and Big Creek passes, 150 feet deep in Markle Pass, and 250 feet deep in Wills Creek Pass.

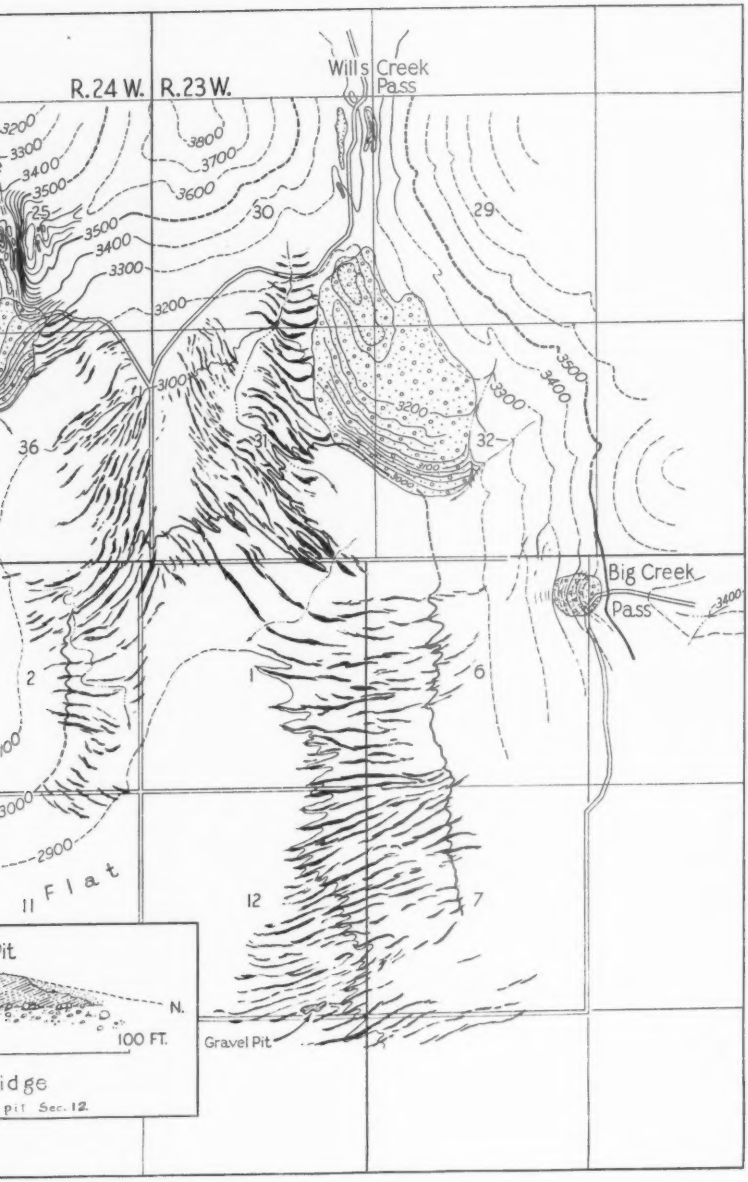
Gap at south.—The south rim of Camas Prairie Basin is lowered by a wide gap to within a few hundred feet of the adjacent basin floor and is penetrated east of the gap by a narrow steep-sided outlet trench. Evidences within the gap of a huge outflowing stream include denuded rock surfaces, discontinuous gravel deposits, and basins, some holding bonds or marshes; all are characteristic of stream beds in rocky channels (Pl. 2, fig. 2). One of the larger gravel deposits (B on Pl. 2, fig. 2) rises from a hollow below an obstructing rock hump to a rounded bar. It extends south to the top of the slope descending to the river opposite Perma where it takes the form of a high eddy deposit, a result, apparently, of the commingling of the stream from the gap with that flowing down the Perma narrows.

Rainbow Lake Pass.—Particularly impressive evidences of stream action appear in Rainbow Lake Pass, a steep-sided trenchlike valley (Pl. 4) which extends, at an elevated position, from the northwest corner of Camas Prairie Basin through the dividing range to Plains Basin on the west.

General Form and Relations: The Camas Prairie side of the pass has two entrances that converge near the mountain axis, about a third of the way across. The northern entrance, which is traversed by the



NORTHERN PART OF CAMAS PR



CAMAS PRAIRIE BASIN

Plains to Elmo automobile highway, is about half a mile wide, and the southern entrance is somewhat narrower. Both commence at an altitude of about 3650 feet at the heads of small valleys that ascend gradually from the floor of the basin some 700 feet below. The pass is about 6 miles long. Its profile descends westward a total of about 50 feet most of which takes place in the branches just above the junction.

Rainbow Lake (Pl. 6, fig. 1) occupies a depression in a narrowed stretch of the pass just below the junction of the two branches. Allowing for small seasonal fluctuations its surface averages about 3600 feet in altitude as determined by aneroid. The lake is fed by two small streams, has no visible outlet, and the water remains fresh.

The pass widens from about half a mile at the lake to nearly a mile where it comes out of the mountains to Plains basin at the west. A view of it from points within the basin, however, is obstructed by a prominent elevation called Locust Hill standing directly in front of the entrance. From Locust Hill a low foothill ridge, parallel to the mountains east of the basin, extends south to the Clark Fork River at Plains (Fig. 4). Between this ridge and the mountains is the relatively narrow valley of Boyer Creek.

South of Locust Hill shelflike terraces are perched along both sides of the foothill ridge within 100 feet of the top. They are about 3550 feet in altitude, 1½ miles long, and toward the south, where the dividing ridge becomes lower, they approach a gap and meet to form a narrow junction. The terrace (Pl. 4) on the western side of the ridge overlooks Plains Basin from a height of 1000 feet. At its southern end the terrace on the eastern side of the ridge is built out about half way across the valley of Boyer Creek and stands about 500 feet above the stream bed. Northward it widens until it extends across the valley. Together these terraces compose Boyer Bench. The Plains to Elmo automobile highway goes up Boyer Creek and gradually ascends the eastern front of the terrace.

Channel Erosion: Rainbow Lake Pass exhibits strikingly the effects of a former large and rapid westward-flowing current. Above the lake the floors of both branches have been severely scoured, exposing the bedrock which remains bare except for a thin rubble of its own fragments loosened by Recent weathering (Pl. 6, fig. 2). The floor is furrowed with channel depressions, several of them holding ponds or marshes (Pl. 4). Upstream angles of prominent knobs are slightly rounded as by water. To a height of 200 feet or more prominent surfaces on the side slopes are bare, though farther up similar surfaces are smoothly covered. The depression occupied by Rainbow Lake re-

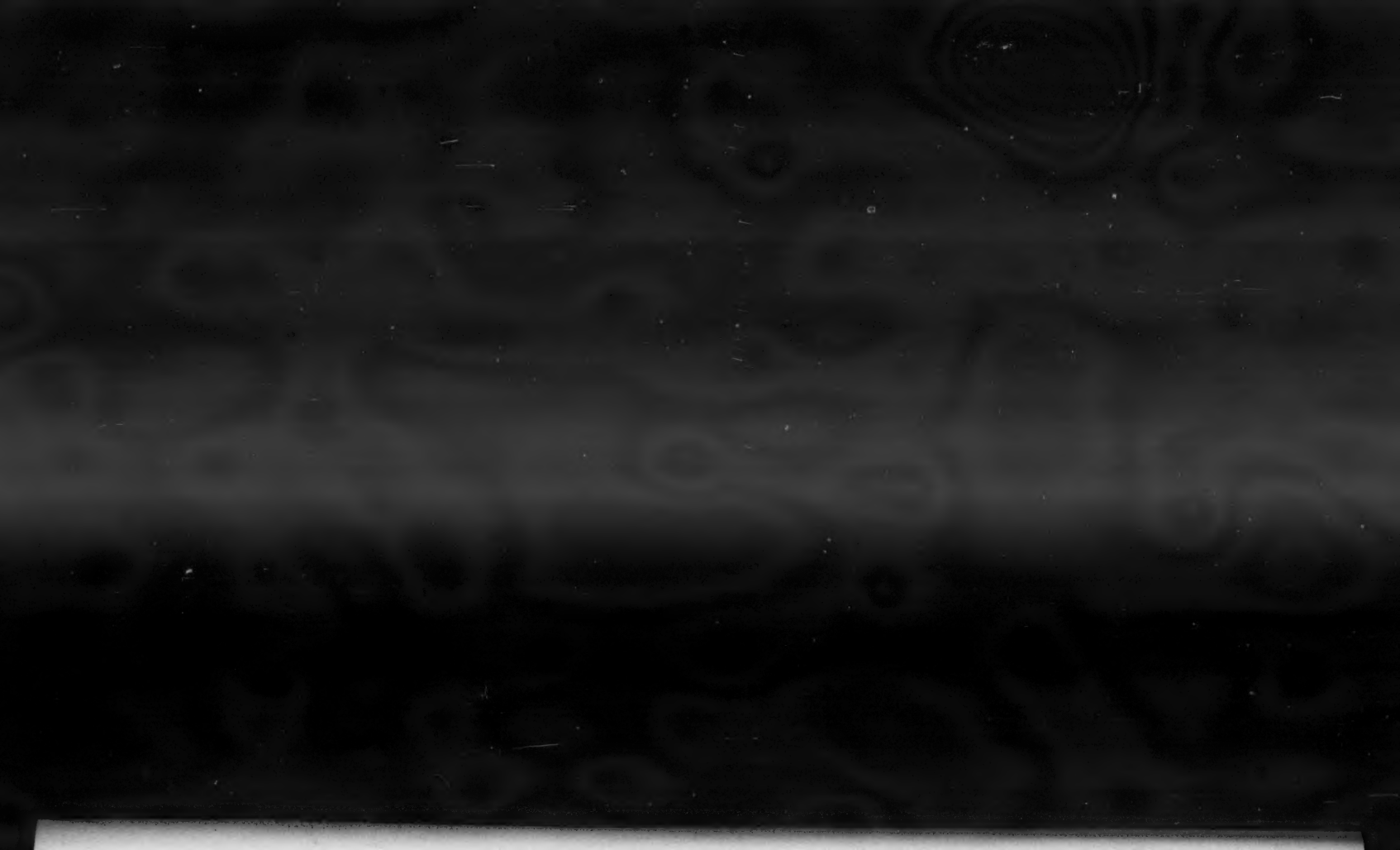
sembles a hollow or "plunge pool" that would be excavated below a rapids by the combined currents from the two branches.

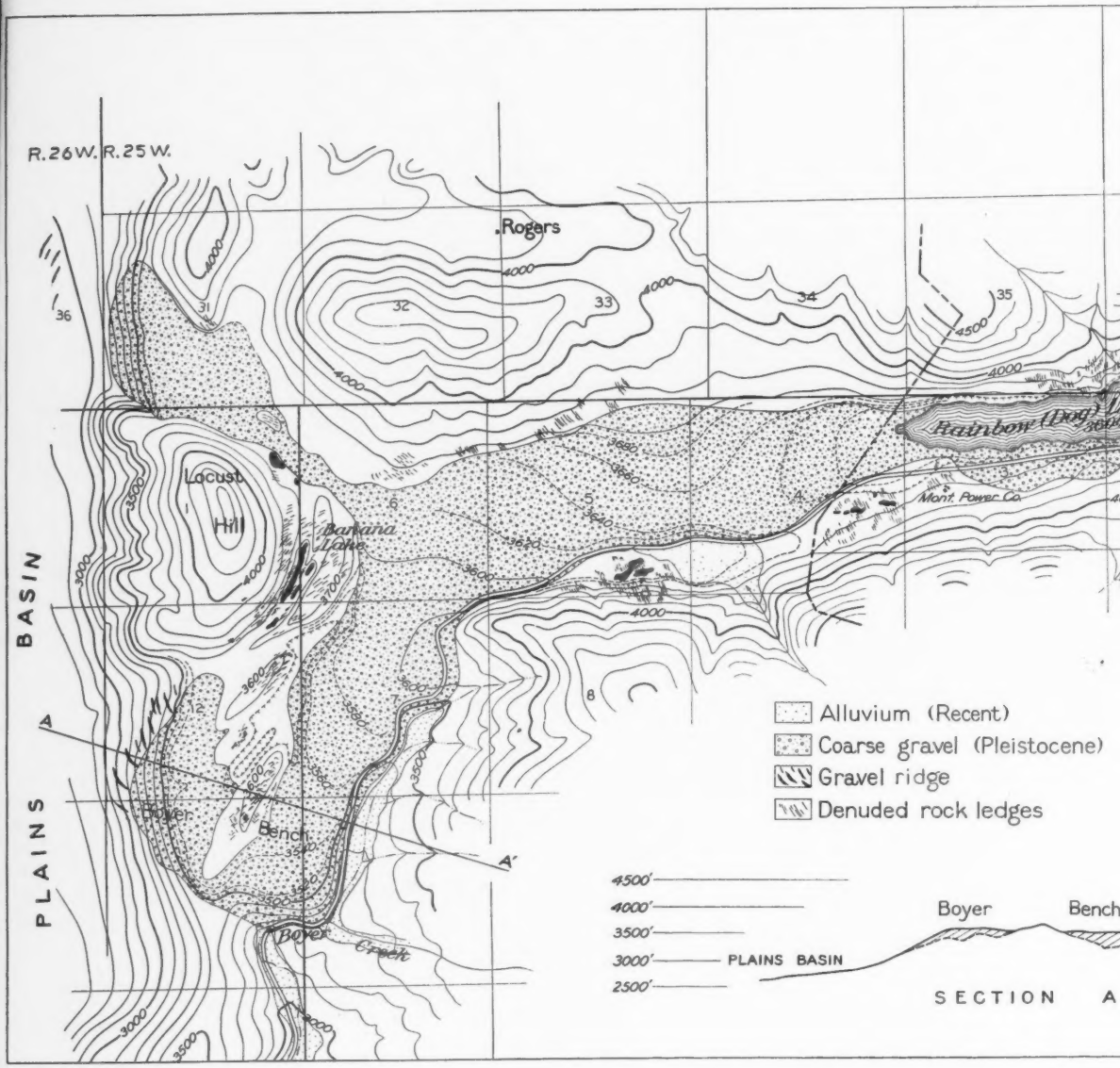
GRAVEL DEPOSIT: Below the lake and along its south side a deposit of coarse stony gravel covers the floor of the pass. Beyond the pass this deposit fills the space to Locust Hill, where it divides into a smaller branch—extending through a gap at the north and expanding in a terrace at the edge of Plains Basin—and a larger branch that turns south into Boyer Creek Valley to form the eastward-facing terrace of Boyer Bench. The entire deposit is poorly sorted, slightly to moderately water-worn pebbles, cobbles, and boulders, most of them between 2 inches and 1 foot in diameter with a variable but generally scanty proportion of smaller pebbles, sand, and silt but no clay. The deposit is not cemented and but slightly compacted. Slightly water-worn to angular fragments, some of them 5 or 6 feet in length, are widely and irregularly scattered through it. A layer 2 or 3 feet thick of "dry gravel," an aggregate of loose stones without sand or other fine matrix, is spread over much of the surface. Included in this layer are fragments of rounded cobbles presumably split apart by frost action since they came to place.

Recent highway excavations along the east front of Boyer Bench expose below the "dry gravel" or other superficial layers a more or less distinct downslope or fore-set bedding (Pl. 8, fig. 1). In some places the beds are lenslike and composed of relatively small pebbles and cobbles in an abundant sandy matrix. In other places the beds are irregular, relatively coarse, and almost without finer material, thus largely consisting of voids. Except for an occasional erratic, the gravel consists of rocks like the outcrops in Rainbow Lake Pass. Most of them are sandstone and argillite of the Prichard formation and Ravalli group which, because of a slaty parting, tend to separate into slabs. A few are blocky fragments of diorite.

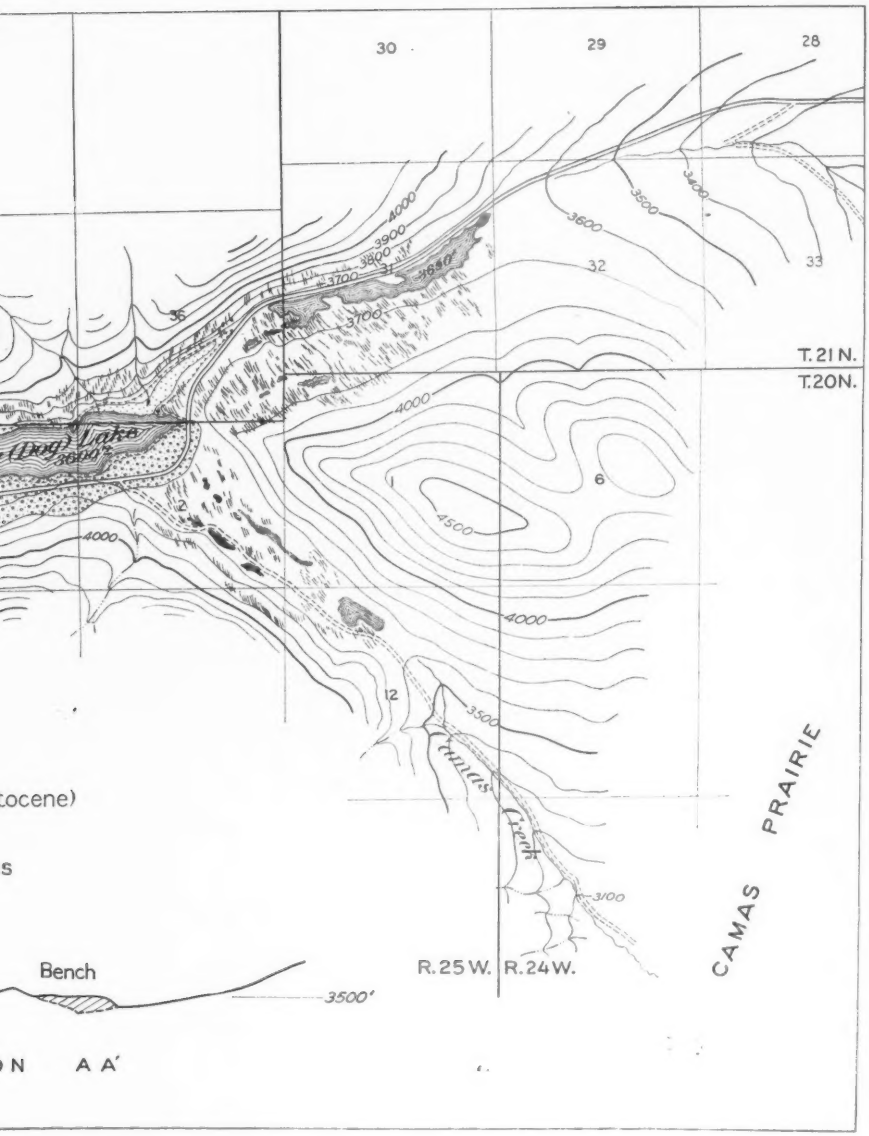
Throughout its area the gravel deposit exhibits the surface characteristics of alluvial stream beds. Of these the minor features include ridges, channels, and basins mostly 5 to 10 feet high or deep but which locally reach 40 feet. The larger features are barlike forms. One of these rises gradually from the lower end of the lake to a height of 70 or 80 feet (Pl. 4). The bar is banked along the north side of the pass opposite a denuded slope on the south side that rises from a channel depression, partly filled with Recent alluvium. Another bar is heaped against the south spur end at the entrance to the pass, a position opposite the scoured east side of Locust Hill.

The ridge at the south of Locust Hill gives evidence of having been overflowed by a large stream from the east. Material eroded from it





RAINBOW LAKE PASS


30
29
28
32
33
35
T. 21 N.
T. 20 N.
4000
3900
3800
3700
3600
3500
3400
4500
(Dog) Lake
6
1
2
12
Phillis Creek
CAMAS PRAIRIE
R. 25 W.
R. 24 W.
Bench
3500'
A A'
2 MILES

forms the western part of Boyer Bench. The surface of this deposit is somewhat roughened with small channels and basins except along the front where it is smoothed by a deposit of later beach gravel.

These features apparently were formed by a huge current swinging from side to side as it moved swiftly through the pass eroding the outsides of bends and depositing gravel bars on the insides. Beyond the pass the stream severely scoured the opposing slope of Locust Hill. There it divided and a smaller part passing around to the north formed an embankment at the edge of Plains Basin. The larger part turned south into the valley of Boyer Creek, deposited the east-facing terrace of Boyer Bench, and, at the same time, overflowed the dividing ridge and formed the terrace facing Plains Basin. The flow was of brief duration, and both branches followed the tendency of a stream passing into deep water to build out a deltaic bar as the channel bed falls away. Presumably the flow started when the surface of the arm of the glacial lake in Plains Basin fell below the surface of the arm in Camas Prairie Basin. It ceased when equilibrium was restored between the two basins or when the water surface in both fell below the pass.

Giant ripple marks of Camas Prairie.—Camas Prairie is distinguished by a number of unique ridges that have the form, structure, and arrangement of ordinary ripple marks but are so large that the term "ripple mark" seems inappropriate. It is retained, however, for want of a better one. Except for a few inconspicuous ones below Big Creek Pass, the ridges (Pl. 3) are grouped in belts appended to the other three northern gaps—namely, Duck Pond, Markle, and Wills Creek passes, a relationship suggesting formation by currents entering the basin through those channels. The relative scarcity of similar ridges below Big Creek Pass might be explained by the relatively small size of its stream.

The distribution, linear form, and arrangement of the ripple-mark ridges as mapped were obtained from aerial photographs (by the U. S. Forest Service) of the belt below Duck Pond Pass, and by pace and compass traverses of the belts below Markle and Wills Creek passes. An unmapped part of the Wills Creek Pass belt extends 1 mile or more farther south. The ridges range from 1 foot or less to 50 feet in amplitude and from a few feet to as much as 500 feet in wave length, but the mean height is between 15 and 30 feet and the wave length about 250 feet. Most of the smaller ridges occur toward the lower (basinward) ends of the belts, and the exceptionally high ones near the middle. Toward the lower end of the belt in front of Duck Pond Pass the ridges gradually decrease in size until they fade out completely.

Individual ridges are from a few rods to half a mile or more in length. Their axes are slightly curved, and convex forward (toward the basin). Their crests are rather wide and rounded and with few exceptions their cross profiles are asymmetrical, the southern or downstream sides being the steeper. As exposed in a pit in one of the ridges about 3½ miles south of Wills Creek Pass (near the southeast corner of sec. 12, T. 19 N., R. 24 W.) the gravel is largely subrounded pebbles ranging from 8 to 20 millimeters in diameter. Sandy matrix is rather scant, the mass is loosely compacted, distinctly stratified, and throughout the cross section the dip of the bedding is parallel to the steeper south (downstream) side (Pl. 3). Toward the pass the gravel of the ridges coarsens, but no other exposures of the structure were seen. In comparable situations the ridges in front of Markle and Duck Pond passes are of similar composition and texture.

The giant ripple marks lie on a smooth bedrock surface that rises gently from the interior flat to the bordering mountains, and, according to the direction which they happen to take across this surface, their profiles are sloping or horizontal. Toward the passes they are overlain by the bar-like deposit of coarse gravel at the end of each channel. All appear to have been formed by large currents plunging over the north rim and continuing with decreasing velocity out into the basin. Such currents might have formed when the water surface in Camas Prairie Basin first became lower than the water surface in Little Bitterroot Basin thus causing masses of water to pass over the ridge, chiefly through the gaps. The initial direction and momentum of the descending masses of water plus increased specific gravity due to the fine sediment picked up on the way would tend to hold the currents to the bottom for a considerable distance.

Giant ripple marks in plains basin.—Gravel ridges that resemble the giant ripple marks of Camas Prairie and are probably similar in origin appear in two places at the eastern side of Plains Basin below Rainbow Lake Pass (Pl. 4, Sec. 12, T. 20 N., R. 26 W., Sec. 36, T. 21 N., R. 26 W.). They are relatively inconspicuous and were first observed in the aerial protographs. The larger ones (Sec. 12, T. 20 N., R. 26 W.) form a group in the path of the stream which overflowed the ridge south of Locust Hill. They are 10 to 20 feet high, 200 to 500 feet apart, and their axes are normal to the course of a current descending to Plains Basin. A group of lower and less definite ridges (Sec. 36 T. 21 N., R. 26 W.) bear a similar relation to the stream that reached Plains Basin north of Locust Hill.

Stream-denuded surfaces in narrows.—Mention has been made of rock surfaces in the gaps around Camas Prairie Basin, in particular Rainbow Lake Pass, that appear as if washed bare of waste mantle except for

FIGURE 1. BEACHES OF GLACIAL LAKE MISSOULA
Northeast side of Missoula Valley at O'Keefe Creek. Hill is about 1000 feet high, summit is slightly above 4150-foot beach which marks highest lake stage. Aerial photo by U. S. Forest Service.

FIGURE 2. PARADISE NARROWS
High eddy deposit with pond at back outlined at B. Light rectilinear streaks are trenches along major joints or shear zones and partly filled with Recent talus. Aerial photo by U. S. Forest Service.

BEACHES AND HIGH EDDY DEPOSIT

FIGURE 1. RAINBOW LAKE
Fills depression in floor of the pass.

FIGURE 2. BEDROCK FLOOR
Surface denuded of waste. Joints opened by weathering.

RAINBOW LAKE PASS

Recent rubble. A similar condition on a much greater scale appears in the narrows along the Clark Fork and the Flathead rivers through which the ponded waters of the Missoula and the Mission groups of basins escaped. In these sluiceways all humps, knobs, and other convexities of the bedrock that project through the sediments of the valley floor or stand out on the adjoining slopes are manifestly bare of a normal waste mantle to heights varying from several hundred to a thousand feet or more. In contrast, comparable slopes at higher altitudes are smoothly covered, as a rule, to a depth of several feet. Thus in the Alberton Narrows below the Missoula group of basins, prominent surfaces on the side slopes are bare several hundred feet up. Knobs and ridges that project above the present flood plain including rock benches 200 feet high on the south side of the river at the entrance to the narrows, and also just below Alberton, are largely denuded of residual mantle or soil. Farther down, in Quinn's Narrows, similar conditions exist.

To the north in the valley of Flathead River denuded rock surfaces begin to show conspicuously along McDonald Narrows, the first noteworthy constriction of the valley below the Mission group of basins. There a spur projects from the north side of the valley, bare of waste to a height estimated at 1000 feet, the slope above being normally covered (Pl. 7, fig. 3). At Perma, 6 miles farther down, the surface mantle is scoured from a rock bench 400 to 600 feet high (Pl. 2, fig. 3) exposing the small scale irregularities of the rock surface as well as several larger trenches and a basin that holds a pond (Pl. 2, fig. 2).

Continuing down through the Perma and Paradise narrows into Plains Basin and on through the Eddy Narrows (Pl. 1, fig. 3) prominent rock surfaces are likewise denuded to a height of 1000 feet or more above the river.

Along the somewhat wider stretches of Clark Fork Valley between the Alberton Narrows and Quinn's Narrows similar conditions are less noticeable, but even there a normal waste mantle is generally absent from the more prominent knobs and spurs. The moderately wide stretch of valley from Thompson River down to the Idaho State line has not been examined in detail, but the lower slopes of Tuscor Hill and other projecting spurs are noticeably bare.

High eddy deposits.—Sediment from confined turbulent streams active enough to keep their channels clear is commonly deposited in recesses along the sides by eddying currents. The gravel bodies described below are thought to be of such an origin. They are like the eddy deposits of ordinary streams that are confined to steep-sided channels, but their production calls for streams of unusual magnitude.

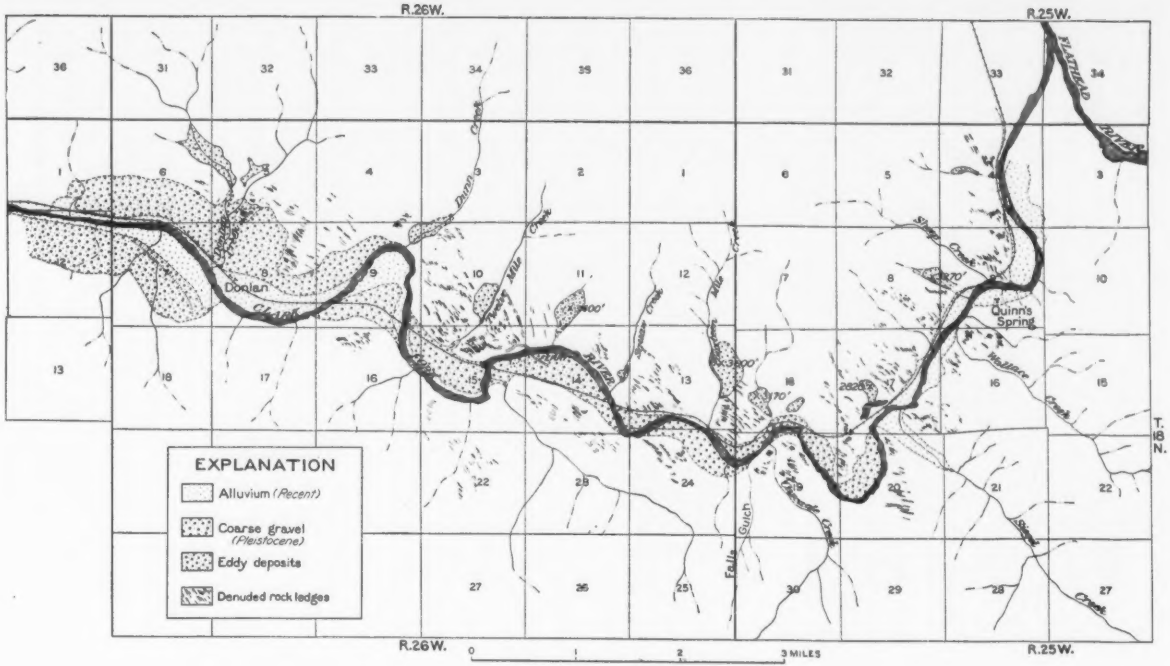


FIGURE 8.—*Quinn's Narrows*

Denuded rock ledges indicated by lines parallel to their trends.

The high eddy deposits occur in side gulches, pockets, or other entrants along the sides of the different narrows. The locations of a number and the heights they attain, as approximately determined by aneroid barometer or triangulation with a Brunton compass, are given in Table 1.

TABLE 1.—*Location and height of High Eddy Deposits in the Clark Fork and Flathead River Valleys*

No.	Location	Height above valley floor (feet)
	PERMA NARROWS (FIG. 5)	
1	Vanderburg Creek (secs. 4, 9)	680
2	E. of Perma (sec. 5)	600
3	W. of Perma (N. ½ sec. 1)	700
4	W. of Perma (center sec. 1)	900
5	W. of Perma (sec. 2)	1000
6	N. of Seepay Creek (sec. 3)	660
7	Burgess Creek (secs. 8, 9)	320
8	E. of Twin Gulch (sec. 4)	300
9	Twin Gulch (Stouts Bar) (secs. 4, 5)	820
10	Robertson Creek (sec. 7)	600
11	E. of Knowles (secs. 5, 6)	840
12	E. of Knowles (sec. 6)	1100
13	NW. of Knowles (sec. 36)	856
14	W. of Knowles (sec. 35)	1200
	PARADISE NARROWS (FIG. 6)	
15	McLaughlin Creek (sec. 21)	610
16	Paradise Gulch (sec. 17)	920
17	W. of Paradise Gulch (sec. 8)	970
18	E. of Henry Creek (secs. 6, 7)	1120
19	Henry Creek (secs. 1, 36)	630
	QUINN'S NARROWS (FIG. 8)	
20	Opposite Quinn's Springs (secs. 8, 9)	790
21	Opposite Siegel Creek (sec. 17)	320

TABLE 1.—*Location and height of High Eddy Deposits in the Clark Fork and Flathead River Valleys—Continued*

No.	Location	Height above valley floor (feet)
22	Opposite Cascade Creek (sec. 18)	575
23	Fourteen Mile Creek (sec. 13)	1035
24	Opposite CCC Ferry (sec. 11)	855
25	Twelve Mile Creek (sec. 10)	645
26	Dunn Creek (secs. 9, 10)	325
27	Donlan Creek (secs. 5, 6)	625
	EDDY NARROWS	
28	S. side below Weeksville	750
29	Poacher Gulch	650
30	Quartz Creek	1000
31	Malone Creek	675
32	Under Kookoosint Ridge	1075

Similar deposits are present in several recesses in Boyer Creek Valley (Fig. 4) along the east side of Plains Basin below Rainbow Lake Pass. More detailed examinations of the different narrows, particularly the more thickly timbered sides, would doubtless reveal others.

The high eddy deposits are characterized by a steep front slope, a rounded top, and a level or gentle back slope as illustrated, for example, by the deposit (No. 16, Table 1) in Paradise Gulch. This body, set somewhat back of the spur end cliffs, begins about 300 feet above the river and rises at a 30-degree slope 600 feet more to a wide rounded top at an altitude of 3400 feet (aneroid). Toward the back its surface descends gradually 100 feet or so to an undrained depression containing a pond (Pl. 5, fig. 2). Back of the depression the buried gulch channel reappears. Another body (No. 18) in the second gulch to the west reaches an extreme height of 1120 feet (about 3500 feet altitude). The profile of its top rises gradually to the west. At the back is a small flat. Still another deposit (No. 9) that is very conspicuous (Pl. 7, fig. 1) from the highway in the valley fills out the lower part of Twin Gulch even with the valley side, its rounded top being about 800 feet above the river. The back slope descends to an alluvial flat drained by a trench at the east.

High eddy deposits formed in gulches with active streams have been more or less washed away. One in the valley of Henry Creek (No. 19) has been separated into two parts, and another (No. 12) in a gulch east

FIGURE 1. STOUT'S BAR
In Twin Gulch on north side of Perma Narrows 2 miles east of Knowles (Fig. 5). Top rises gradually toward left. Maximum height above river about 800 feet. Adjoining rock slopes on right and left cleanly scoured of waste.

FIGURE 2. NORTH SIDE OF QUINN'S NARROWS OPPOSITE SIEGEL CREEK.

FIGURE 3. NORTH SIDE OF MCDONALD NARROWS
Denuded rock spur with deposit in shallow depression. Varved silt forms terrace at foot of slope.

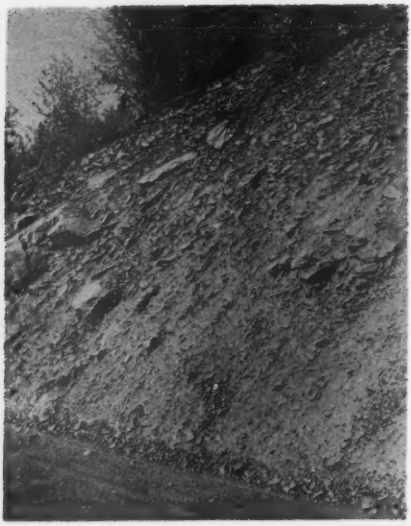


FIGURE 1. GRAVEL OF BOYER BENCH
Road grade up east front of terrace.

FIGURE 2. BEACHES ON WEST SIDE OF BOYER BENCH
Uppermost beaches turn into gap at right and end in small spitlike embayments.

FIGURE 3. GRAVEL WITH FRAGMENTS OF VARVED SILT
Pit (Fig. 4, Sec. 9, T. 20 N., R. 26 W.) east of Clark Fork River near entrance to Eddy Narrows.

of Knowles has been extensively cut away on the east side. In contrast, deposits that have been protected from stream cutting remain unmodified as illustrated by one in Quinn's Narrows (No. 21) that occupies a niche (Pl. 7, fig. 2) in the valley wall opposite Siegel Creek.

The gravel of the high eddy deposits consists of partly rounded pebbles mostly less than 2 inches in diameter in an abundant sandy matrix. In places the gravel is coarser, and a few large rough slabs like the rock of adjoining ledges usually are scattered through the mass. Gravel pits in deposits on Dunn Creek (No. 26) and on the west side of Boyer Creek Valley just above Plains expose distinct fore-set bedding.

At first glance the high eddy deposits suggest such features as deltas, bay-mouth bars, or terrace remnants of an alluvial valley fill. Attempts, however, to explain them as such meet with difficulties. Their back slopes descend in the opposite direction to the rising surface of a delta, and many of the larger ones are in pockets or disproportionately short gulches. Although a lagoonlike depression is generally present at the back the blockading ridge, unlike the horizontal top of a bay-mouth bar, has a sloping or rounded profile. The deposits, though situated in narrows where wave action was too weak to develop noticeable strand lines, are generally larger than the beaches and bars that were formed by the relatively strong waves in the wider basins. In addition their vertical distribution is without relation to any definite still stands of the lake. An interpretation of the deposits as remnants of a gravel fill of the river valley seems definitely disproved by their forms, vertical distribution, and the lack of correlative remnants of such an extensive deposit outside the narrows. On the other hand their diminutive counterparts may be observed in almost any rough-sided ditch or gutter after a flood has subsided.

Coarse gravel deposits in Clark Fork Valley.—Certain deposits of coarse bouldery gravel in Clark Fork Valley apparently are remnants of more extensive bodies laid down by a huge and rapid stream. One of these partly covered by sand and silt occupies a large area in the western part of Plains Basin. Its cross profile is slightly convex and along the east it has been partly cut away by the river from which it rises as a terrace 100 feet or more high (Pl. 1, fig. 1). Westward it extends to the side of the valley causing Swamp Creek to turn abruptly northward and hug the foot of the mountains (Fig. 4). The form and situation of this body suggests a channel deposit of a huge stream projected from Paradise Narrows, spreading and slackening somewhat as it crossed the basin to the outlet, the Eddy Narrows.

Where it enters the Clark Fork Valley just below the Eddy Narrows, the Thompson River has cut through a superficial layer of comparatively fine gravel (Fig. 2) and exposed a mass of large boulders over which it descends in a rapids. The position of the boulder mass suggests deposition by a huge stream rushing out of the Eddy Narrows.

About 4 miles below the Eddy Narrows a deposit of coarse gravel (Fig. 2), partly concealed by a superficial cover of fine sand, occupies a large area in the Clark Fork Valley adjoining Thompson Falls. From a terrace front facing the river it rises very gradually into a rounded barlike form that extends to the northeast side of the valley and turns the drainage from the mountains to a northward course along its margin. The form, position, and extent of this mass suggest a bar deposited by a stream large enough to have occupied the full width of the valley. Apparently the current turned against the southwest side from which it was deflected across the valley by a precipitous wall (Flatiron Ridge) below Thompson Falls, thus completing a wide bend. The current eroded the outside of the bend and deposited a bar on the inside. In the vicinity of Thompson Falls several large angular rocks lie on or in the gravel deposit. These may have been detached from ledges that crop out just above the town or transported from a distance by floating ice.

AREA AND VOLUME OF THE GLACIAL LAKE

The maximum altitude of the water surface in Glacial Lake Missoula is placed at 4200 feet as indicated by the uppermost of a series of beaches (Pardee, 1910). Repeated checks with the aneroid have indicated that the altitude of the highest definite beach is 4150 feet, although some doubtful beach gravel was found in one place (divide east of Oliver Gulch in T. 21 N., R. 22 W.) near the 4200-foot level. The outline of the lake at the 4150-foot level (Fig. 3) has been prepared with the aid of topographic maps and other surveys that were made after 1910. As thus determined, the area of the lake in round figures was 2900 square miles, and its volume 500 cubic miles. The areas, average depths, and volumes of water held by the different arms of the lake are roughly estimated (Table 2).

DEVELOPMENT OF THE UNUSUAL CURRENTS

Apparently the ice dam failed sufficiently to permit a sudden great outrush of water. The arm of the lake just above the dam was affected first, and there being no obstacle to its free discharge its surface fell rapidly throughout. Beyond Thompson River, however, the outflow was restricted by the Eddy Narrows and successively held back by the

several narrows farther up. The lowered water surface in the arm above the dam induced a strong current through Eddy Narrows which lowered the water surface in the Plains Arm. As a result inflows were caused through Rainbow Lake Pass from Camas Prairie Arm and through the

TABLE 2.—*Approximate areas, depths, and volumes of arms of Glacial Lake Missoula*
The water surface was at 4150 feet altitude.

Arm	Area (Square miles)	Average depth (Feet)	Volume (Cubic miles)
CLARK FORK VALLEY FROM MONTANA-IDAHO STATE LINE TO MOUTH OF FLATHEAD RIVER			
Lower Clark Fork (State line to Eddy Narrows)	465	1400	123.3
Thompson River Valley	115	600	13.0
Eddy Narrows	20	1380	5.2
Plains Basin	90	1350	23.0
Paradise Narrows	10	1330	2.5
FLATHEAD RIVER VALLEY			
Perma Narrows	10	1280	2.4
Camas Prairie Basin	95	870	15.6
McDonald Narrows	25	1200	5.7
Little Bitterroot Basin	260	900	44.3
Jocko Basin	105	900	17.9
Mission Basin	575	1100	120.0
CLARK FORK VALLEY ABOVE MOUTH OF FLATHEAD RIVER			
Quinn's Narrows	35	1200	8.0
Saint Regis River Valley	175	600	18.2
Superior Arm (Quinn's Narrows to Alberton Narrows)	120	940	21.4
Alberton Narrows	30	900	5.1
Missoula-Ninemile Basin	245	740	34.3
Bitterroot Basin	405	700	53.7
Hellgate Arm (Clark Fork Valley above Missoula Basin)	50	150	1.4
Blackfoot River Valley	65	440	5.4
Totals	2895 square miles		520.4 cubic miles

Paradise and Perma Narrows from that and other arms farther up (Fig. 9). The consequent lowering of the water surface in the Camas Prairie Arm caused an inflow across the submerged northern rim. In succession the other narrows released outflows from the more distant basins. The subsequent behavior of the currents thus generated was controlled largely by such topographic relations as the distribution and capacities of the different basins, the lengths and cross sections of the narrows, and in particular the "bottle necks" formed by the Paradise and Eddy narrows.

VOLUME AND DURATION OF THE UNUSUAL CURRENTS

Evidently the streams were of unusual size but were short-lived because of limited water supply. No precise data, of course, are available. The

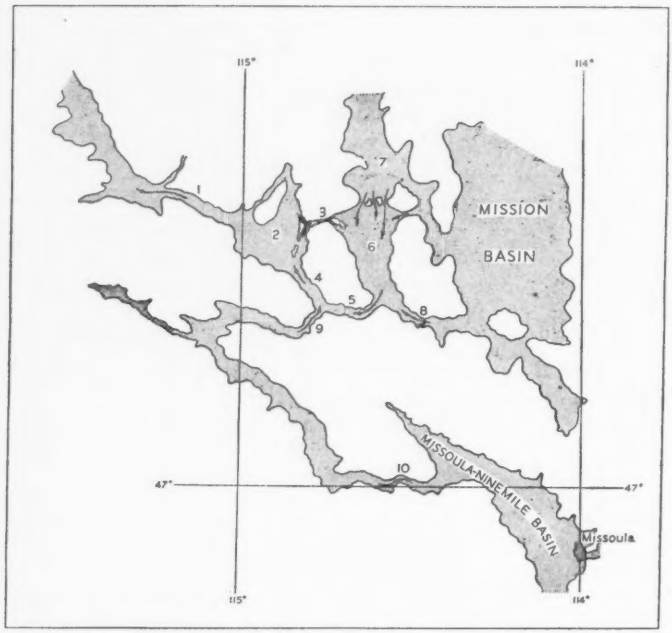


FIGURE 9.—*Central part of Glacial Lake Missoula*

Medium stage.

1. Eddy Narrows.
2. Plains Basin.
3. Rainbow Lake Pass.
4. Paradise Narrows.
5. Perma Narrows.
6. Camas Prairie Basin.
7. Little Bitterroot Basin.
8. McDonald Narrows.
9. Quinn's Narrows.
10. Alberton Narrows.

following results, based on formulas for small streams, are introduced merely to give some idea of their size.

Eddy Narrows provided a nearly straight sluiceway 10 miles long with no very prominent projections along the sides to hinder the flow. As indicated by its high eddy deposits and the denudation of its sides the stream was at least 1000 feet deep. Available information of the upper limit is not precise enough to permit a satisfactory determination of the surface gradient, which is therefore assumed to have been the same as the average

descent of the present river bed—namely, 2.3 feet per mile through the 10 mile stretch of the narrows. The average cross section is about 5,900,000 square feet, and the wetted perimeter 9600 feet. Using these figures with the Chezy formula (Merriman, 1931) for rough stone channels gives a velocity of 52.3 feet per second. According to the formula adopted by Manning (Lindquist, 1934) the same figures would indicate a velocity of 78 feet per second. The velocity of the eddying currents is rather closely related to that of the main stream (Leighly, 1934). As determined by a maximum pebble size in general of 3 inches (76 mm. approximately) the currents that formed the high eddy deposits had (Hjulstrom, 1935) a velocity of about 65 feet per second which may be assumed as about the same as the main current. Incidentally, this figure happens to be the mean of the two other determinations. Accepting it as approximately correct the flow through the Eddy Narrows would appear to have reached a peak of 386,500,000 second feet or about 9.46 cubic miles per hour. As only about 380 cubic miles of water was impounded above the Eddy Narrows the maximum flow could not have continued for long. The maximum discharge of the Mississippi River at Natchez in the flood of February 1937 (Williams and Crawford, 1939) was 2,100,000 second feet or about .05 cubic mile per hour.

Sufficient data are not available for even rough estimates of the flows carried by the other narrows or by the wind gaps around Camas Prairie or by Clark Fork Valley near the ice dam. However, according to data compiled by Hjulstrom (1935) the velocity of the stream that formed the giant ripple marks is indicated at one place (gravel pit in SE¼ sec. 12, T. 19 N., R. 24 W.) by a maximum pebble size of 40 mm. to have been 7 feet a second. This current presumably was at least twice as deep as the ripple marks are high, or about 100 feet, and it may have occupied the full width of the basin floor; obviously no estimate is possible without more data.

AGE OF THE UNUSUAL CURRENTS

Although they have remained exposed since the water subsided, the giant ripple marks and other features produced by the unusual currents show very little effect of weathering. Rock surfaces stripped of their surface mantle remain bare except for a relatively thin new cover. Talus aprons generally do not reach more than one third to half way up a cliff (Pl. 2, fig. 1). Except for later stream trenching in a few places, the giant ripple marks look as though they had just emerged. Where protected from stream erosion the high eddy deposits are likewise unmodified. Few of the many channel depressions in the wind gaps

around Camas Prairie Basin have become filled. Large parts of the coarse gravel deposits in Plains Basin and Clark Fork Valley at Thompson Falls have survived attacks by the river. Altogether these conditions suggest an age not greater than Middle Pleistocene.

LAST STAGE OF THE GLACIAL LAKE

Either the rapid outflow was greatly slowed when the lake was partly drained or, if it had become emptied, the basin soon thereafter became, partly at least, refilled, and in this new body of water no unusual currents seem to have developed. The evidence for this delayed drainage or refilling consists mainly of beaches formed on the current-produced features of the preceding outflow.

A series of small beaches are plainly visible on the side of Boyer Bench facing Plains Basin (Pl. 8, fig. 2). From rather faint lines below Locust Hill they become more definite southward where the higher ones, as they turn eastward around the terrace, end in small spitlike embankments. Elsewhere rather faint strand lines are visible on a high eddy deposit along Vanderburg Creek east of Perma (NW¼ sec. 9, T. 18 N., R. 23 W., Fig. 5) and on the basinward side of the coarse gravel deposit in front of Wills Creek Pass (Pl. 3). A small gravel bar continuous with a beach at 3400 feet altitude lies across the lowest part of the stream channel of Markle Pass at its northern entrance. Another blocks the north entrance to a channel about 140 feet higher (Fig. 7). A similar relation between beaches and channel features exists at the northern entrance to Wills Creek Pass.

The highest of the later beaches is at 3540 feet. Whether that figure is the upper limit of the last submergence cannot be positively stated. Where the outflow features are absent present knowledge of the beaches is not sufficient to distinguish the younger from the older.

A relatively slow current during the final outflow may have formed the gravel bar in Plains Basin east of the river near the entrance to the Eddy Narrows. As exposed by a pit along the highway (Pl. 8, fig. 3) the gravel is fine-textured (about 15 to 20 mm., pebble size), clean, and its bedding fore-set. The gravel rests on a platform about 50 feet higher than the river and contains fragments of the varved silt.

SUMMARY

Failure of the ice dam in the Clark Fork Valley near the Idaho-Montana State line permitted the rapid draining of Glacial Lake Missoula estimated to have held 500 cubic miles of water. Rapid falling of the water surface caused huge currents in the narrows between sub-

merged basins. Current-produced features include the unique giant ripple marks of Camas Prairie, a series of gravel ridges having the form, structure, and arrangement of ordinary ripple marks but of incomparable size. The flow through Eddy Narrows is estimated to have reached a maximum of nearly 9½ cubic miles an hour. After the outflow the basin again held a lake which drained away slowly.

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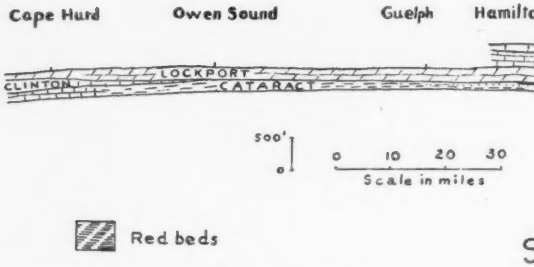
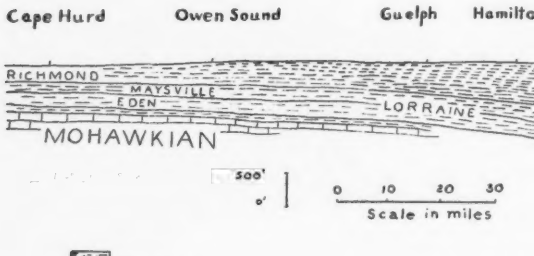
ONTARIO





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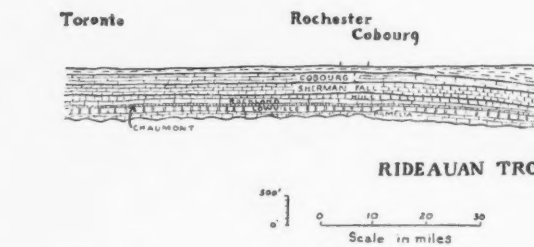

ALLEGHENY BELT

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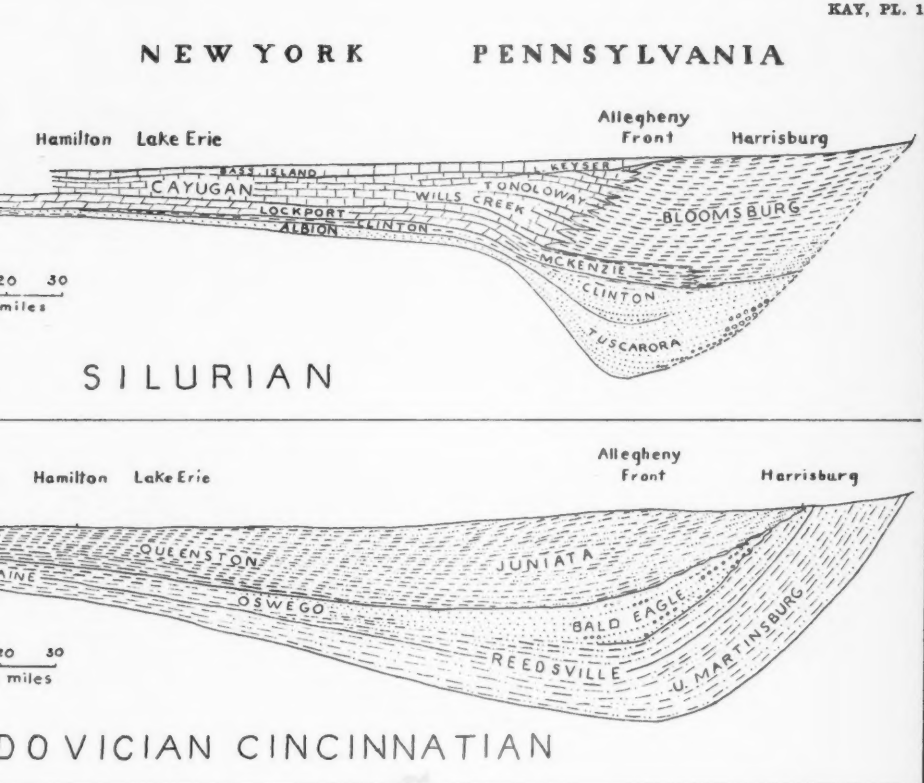

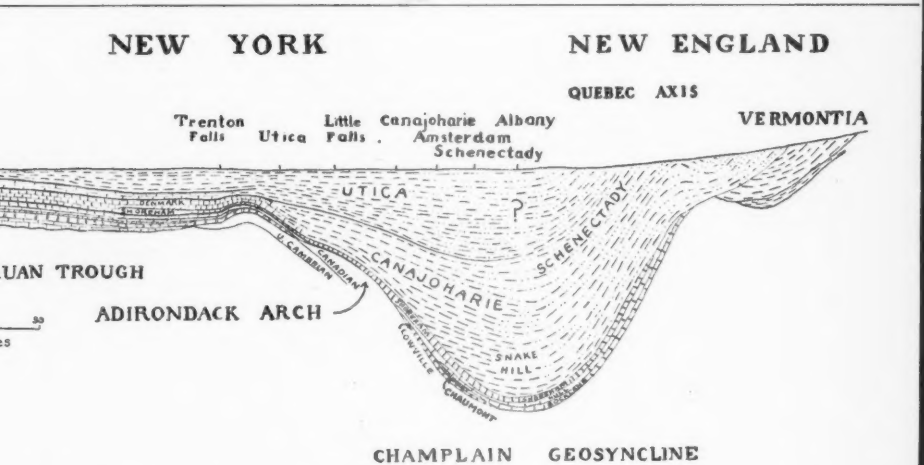

DOVICIAN AND SILURIAN RESTORED SECTIONS

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DEVELOPMENT OF THE NORTHERN ALLEGHENY SYNCLINORIUM AND ADJOINING REGIONS

BY G. MARSHALL KAY

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ABSTRACT

Analysis of the Paleozoic stratigraphy in the northern Allegheny synclinorium indicates that (1) Gross deformation was downward in early and middle Paleozoic. (2) The greatest stratigraphic thicknesses are in extensive semilenses of detrital sediment laid in depressions complementary to uplifts of the Taconian and Acadian orogenies, in deltageosynclines. (3) The Cincinnati and Adirondack axes geographically define the Allegheny belt; they had arches for limited times, were hinges of oscillation, and at other times were more depressed than the central part of the belt. (4) The axes were less depressed than the elliptical Michigan geosyncline to the west and the linear Champlain geosyncline on the east. (5) There may have been warping (epeirogenic) and sea-level (eustatic) movements synchronous with and related to the orogenies, as well as depression produced by loading (isostatic adjustment); the causes and interrelations are not proven, however.

Later Paleozoic structures trend more southwesterly than the Adirondack axis; the rectilinear margin of the late Paleozoic (Virginian) geosynclinal belt crosses the axis in the central Appalachians. The eastern margin of the southern part of the Allegheny belt is concealed beneath the first extensive Appalachian thrust sheet. Inasmuch as this sheet bears middle Ordovician rocks stratigraphically similar to those in the Champlain belt in the north, it forms an allochthone, and the thrust developed along the Adirondack axis. Appalachian folds that extend into the northern Allegheny synclinorium parallel the Appalachian Structural Front, the western principal fold in the mountains. The Front has the arcuate trend of deltageosynclinal isopachs in the north and of the rectilinear trend of the margin of the Virginian belt in the south. The northern margin of the synclinorium was block-faulted in the early Tertiary.

INTRODUCTION AND ACKNOWLEDGMENTS

The Allegheny synclinorium is a broad, northeast-trending structural depression east of the Cincinnati arch (Fig. 1). Its northern margin extends to the faulted pre-Cambrian highlands of central Ontario and the Frontenac axis at the outlet of Lake Ontario. On the east lie the Adirondack block mountains. Farther south, the Appalachian Structural Front separates the deeply folded Appalachian Mountains from the gently warped border of the "geosyncline" beneath the Appalachian Plateau; though the front forms the margin of the present synclinorium, the extension of the Adirondack axis is farther east. The several limiting uplifts differ in character and times of origin, and the synclinorium is complex. Earlier deformation is evidenced in the distribution, thickness, facies, and conformability of the sediments, for much of the depression is of Paleozoic origin. Later history is recorded in the changed attitudes of the deposits.

The present survey has developed from stratigraphic studies of the Ordovician sediments along the northern and eastern margin of the region. Development of the region as a whole has been discussed rarely. Much information on which conclusions are based is in the literature, to which reference will be made continually. Additional data on subsurface have been furnished generously by the State surveys of New York, Michigan, and Ohio, and the Department of Mines of Ontario.

Students have continually encouraged the writer toward reasoned explanations of stratigraphic relations. The publications of Bucher, Grabau, Keith, Moore, Ruedemann, Schuchert, Stille, Ulrich, and Van der Gracht have particularly directed endeavors. W. O. Clif and Betty R. Smith have assisted in the preparation of illustrations.

The typescript was submitted originally in April 1940, as part of a paper, *Development of the northern Allegheny synclinorium and adjoining Canadian Shield;* the complementary part has been published as the *Ottawa—Bonnechere graben and Lake Ontario homocline* (Kay, 1942). The present article was awarded the George Frederick Kunz Prize in Mineralogy and Geology by the New York Academy of Sciences in 1941.

EARLIER PALEOZOIC STRATIGRAPHY

INTRODUCTORY STATEMENT

An understanding of the development of the Allegheny synclinorium must be gained from the stratigraphy and structure of the rocks of the region. The material presented in the following pages is principally descriptive, a summary of information accessible in scattered literature. It forms the basis for the interpretations in the latter part of the paper.

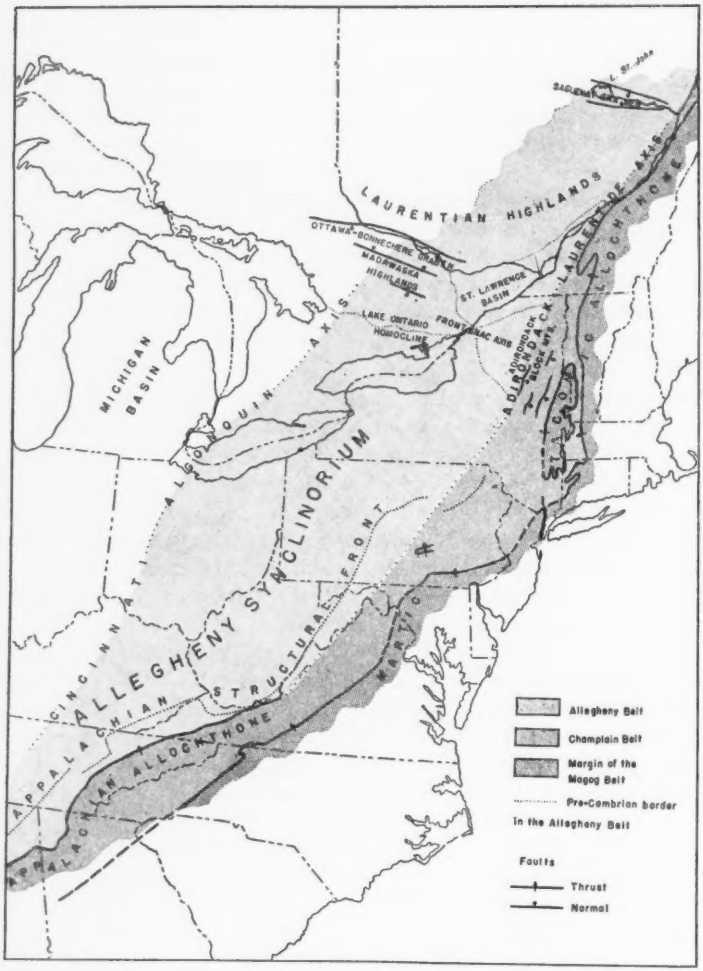


FIGURE 1.—*Map locating the Allegheny synclinorium, structural and stratigraphic provinces*

The limiting structures of the northern Allegheny synclinorium became defined in the Ordovician and Silurian. Throughout the earlier history of the belt, developments in the region to the east were particularly significant in its record. Belt and axis are to be used in a geographic sense without implication of structure. The Allegheny belt lies between the axis of the middle Ordovician Adirondack arch and the early Silurian Cincinnati arch. Although much of it lies within the Allegheny synclinorium, a present structure, the northeast part extends beyond the Frontenac axis, and the southern into the folded Appalachian Mountains and beneath the Appalachian allochthone.

The Cambrian and earlier Ordovician rocks, sparse in the northern part of the synclinorium, are well represented in the Hudson-Champlain lowland east of the Adirondack Mountains. They were laid in the Champlain geosyncline, separated in the early Ordovician by uplifts along the Quebec axis from the Magog geosyncline farther east (Fig. 2); the axial structure directed and limited the westward extension of terrigenous lower Ordovician clastics, so that preponderantly argillaceous rocks of the Magog belt contrast with carbonates in the Champlain. The Adirondack arch came to define the western margin of the Champlain trough in middle Ordovician, separating it from the Allegheny belt.

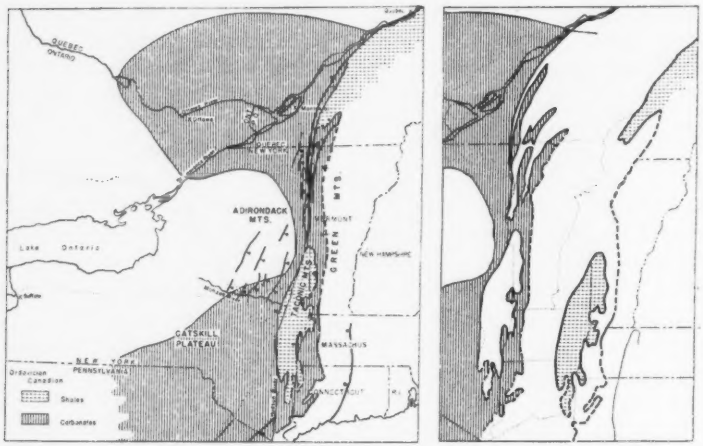
In middle Trenton time, a belt of uplift, Vermontia, formed within the western part of the Magog geosyncline, resulting first in the introduction of terrigenous sediments in the Champlain trough and, in later Ordovician, into the Allegheny belt. At the close of the period, in the Taconian revolution, the argillaceous rocks of the Magog belt were thrust far westward, forming the Taconic thrust sheet that conceals the Quebec axis and all but the western part of the Champlain trough sediments.

The Adirondack and Laurentide Mountains lie along the Adirondack axis, and at least the former was block-faulted, probably at the beginning of the Silurian. Structure is first evident along the Cincinnati axis in early Silurian and continued to develop through the Paleozoic; the Michigan basin to the west also is clearly shown at this time. Lower Devonian Oriskanian sediments are limited northward south of Lake Ontario; the presence of equivalent rocks at Montreal implies a cross structure near the present Frontenac axis. Thus, the northern Allegheny synclinorium was delimited in the Devonian.

CAMBRIAN SYSTEM

The oldest sediments in the northern Allegheny belt are Upper Cambrian (St. Croixan). They overlap older Cambrian rocks of the Lake Champlain region and central Pennsylvania and extend to the vicinity of Utica on the Mohawk River, the Thousand Islands, and Arnprior on

the Ottawa River (Fig. 3a); they are present in Adirondack outliers and northeast of Montreal. Though absent at Quebec City, where Mohawkian overlaps pre-Cambrian, they must be beneath the Taconic thrust plate to the south, for their boulders are present in the allochthonous lower Ordovi-



a. Present distribution of Canadian facies.

b. Canadian facies on palinspastic map.

FIGURE 2.—*Ordovician facies on present and palinspastic maps*

The carbonates in the Champlain belt were separated from the shales in the Magog belt by the Quebec "barrier". After Kay (1937) and Cady (1938).

cian shales on the margin of the plate. Cambrian sediments are absent along the margin of the Paleozoic outcrop between the outlet of Lake Ontario and the head of Lake Huron but reappear on St. Joseph Island near the Michigan boundary. None are in wells penetrating to the pre-Cambrian southward to lakes Ontario and Erie or south of the former in New York. Questionably Cambrian rocks are present at depth in Sandusky County, Ohio, along the Cincinnati axis (Ballard, 1938, p. 1552). In Michigan, Cambrian is absent in the Detroit region and has not been reached by wells in the basin proper.

The margin of Cambrian deposits is sinuous, a northwest-trending trough, the Ottawa Bay, extending from the Champlain geosyncline. Though the limit in the Adirondack region is not exactly determinate because outliers are few and poorly distributed, the data show the presence of a low peninsula lacking apparent correlation with the present mountains; the land Adirondackia (Schuchert, 1910, p. 463; Kay, 1937, p. 289) is prob-

ably misconceived. The westward thinning is best shown along the Mohawk Valley, where about 100 feet of Potsdam sandstone is overlapped westward by about 500 feet of Little Falls dolomite, which disappears in the West Canada Creek valley, probably through overlap and offlap (Ulrich and Cushing, 1910); there is no evident structural influence along the Adirondack axis.

ORDOVICIAN SYSTEM

Canadian and Chazyan series.—The limited distribution of Canadian and Chazyan rocks in the northern Allegheny belt is comparable to that of the Cambrian. At the beginning of the Ordovician, the Ottawa Bay continued to receive sediments, the distribution of which is summarized in Figures 3b and 3c.

Canadian rocks crop out westward to Little Falls on the Mohawk, the Thousand Islands on the St. Lawrence, and Allumette Island on the Ottawa, and trend northeastward from Montreal to north of Three Rivers. They offlap the Cambrian in New York and overlap somewhat in the Ottawa Valley. The thickness increases from a few tens of feet in Renfrew County, Ontario, to more than 2000 feet in the Lake Champlain region. Canadian rocks have not been encountered in wells penetrating to the pre-Cambrian in southwest Ontario or in New York and Ohio immediately south of the Great Lakes. They are present in the subsurface farther south in Ohio and, though absent at the Detroit River, have been found farther west in Lenawee and adjoining counties. A large peninsula occupying most of the Allegheny belt extended beyond the present lakes Ontario, Erie, and Huron.

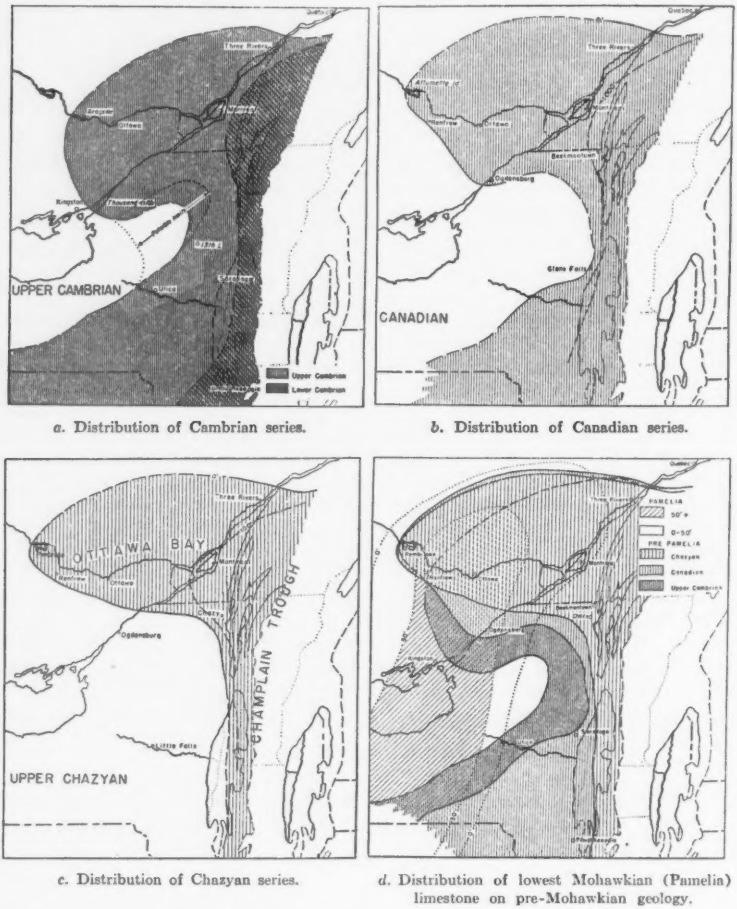
That the Canadian history involved transgression, regression, lesser transgression, and withdrawal is shown by the relatively greater distribution of basal Canadian carbonates and the presence of younger Canadian disconformably on them, as at Ogdensburg, New York (Cushing, 1916).

The distribution of the Chazyan is somewhat similar (Fig. 3c). The overlapping beds thin northwestward from 1000 feet along the west margin of the Champlain trough; the thickness increases eastward in Vermont (Cady, 1938). The Chazyan overlaps in the Ottawa Bay, but the younger beds offlap northward along Lake Champlain (Raymond, 1906, p. 570). The most evident dissimilarity with distributions of older series is the absence of Chazyan south and southwest of the Adirondacks; and, whereas the Canadian and Chazyan are about 4000 feet in the Allegheny belt in central Pennsylvania, they are lacking farther north.

Mohawkian series.—Mohawkian sediments are extensive along the northern margin of the synclinorium and northeastward in the Allegheny belt. Their distribution shows a changed structural pattern from that of the Chazyan (Fig. 3d), evidencing the development of a foreland fold, the Adirondack arch, separating the Champlain trough from the newly formed Rideauan trough [1] in which the oldest Mohawkian of the region (Pamelia) was laid (Fig. 3d); the oldest Trenton beds (Selby member

[1] Rideauan (new) replaces Ontarian (Kay, 1937, p. 289), the Rideau Lakes and River lying along the axis; Ontarian is a name too similar to Ontario basin (Schuchert, 1910, p. 473), applied to a different area, and is a homonym of the pre-Cambrian Ontarian geosyncline (Schuchert and Dunbar, 1933, p. 97).

of the Rockland formation) have similar restriction (Fig. 5a). The Adirondack belt has thin and incomplete sections of the Black River and lower Trenton limestones; their irregular distribution suggests an un-



a. Distribution of Cambrian series.

b. Distribution of Canadian series.

c. Distribution of Chazyan series.

d. Distribution of lowest Mohawkian (Pamelia) limestone on pre-Mohawkian geology.

FIGURE 3.—*Cambrian and Ordovician paleogeography*

dulating surface on which the seas had sinuous and embayed shores. The character of the belt is not conclusively shown, but there were no significant highlands. That it became an arch (Kay, 1936) separating the

Champlain trough from the depressed Allegheny belt is shown in the distribution and character of the Trenton formations.

The more restricted Mohawkian seas were margined on the north by shores trending east-west. The basal Trenton overlaps the Black River along a line extending from the north shore of Lake Huron toward Lake St. John, Quebec; and the higher Sherman Fall is overlapped by the Cobourg along a comparable line.

The Vermontian disturbance of middle Trenton time raised a highland within the western part of the Magog geosyncline and directly influenced the character and thickness of sediments in the Champlain trough, but terrigenous sediments entered the Allegheny belt only in latest Trenton (Collingwood). The disturbance is further reflected in the Allegheny belt only if the late middle Trenton marine withdrawal resulted from related eustatic fall or warping.

Mohawkian sediments overlap crystalline rocks over large areas (Fig. 3d), the Trenton extending beyond the Black River north of the present synclinorium. The oldest Black River (Pamelia) limestone evidences the southwestward trend of the axis of the Rideauan trough, being thickest at the northeast end of Lake Ontario and near Ottawa. The formation thins and overlaps toward Georgian Bay, is about 150 feet at the St. Lawrence River, and absent southeast of the Black River Valley. Along a parallel line northeast, it is absent at Mattawa, approaches 100 feet west of Ottawa (A. E. Wilson, 1934), thins eastward to Montreal (Okulitch, 1936) and northeastward (Okulitch, 1939) along the southeastern margin of the Allegheny belt. Pamelia sediments are lacking in the Lake Champlain region.

The Lowville is more extensive and is about 50 feet thick, north and east of Lake Ontario. It thins to extinction along the Mohawk toward the Adirondack axis and is relatively thin at Montreal and northeastward. It is present west of Lake Champlain but disappears east of the lake (Cady, 1938). The Lowville is overlapped by the Chaumont around pre-Cambrian inliers north of Lake Ontario and in exposures north of Lake Huron and in Lake Nipissing. Chaumont limestone is normally about 30 feet thick along Lake Ontario, thins northeastward, and is less extensive than the Lowville along the Mohawk and northern Lake Champlain. Thickness is normal at the southern end of the lake (Kay, 1937, p. 259), and Chaumont-like beds crop out in the outlier at Wells in the Adirondacks. Black River beds are absent at Quebec City and along Lake St. John (Raymond, 1916). Young (1937) has studied the stratigraphy of northwestern New York and southeastern Ontario.

In summary, the oldest Black River (Pamelia) is thickest in the Rideauan trough, thinning by overlap and convergence toward its margins. The Lowville is more extensive and in turn is overlapped along the northern margin of the trough by the Chaumont. Although the distribution in the Adirondack belt might be attributed to a breached arch, it is probable that the Lowville and Chaumont seas extended eastward as far as Vermont with embayed shore lines and that the beds had been somewhat eroded from the Adirondack axis in pre-middle Trenton.

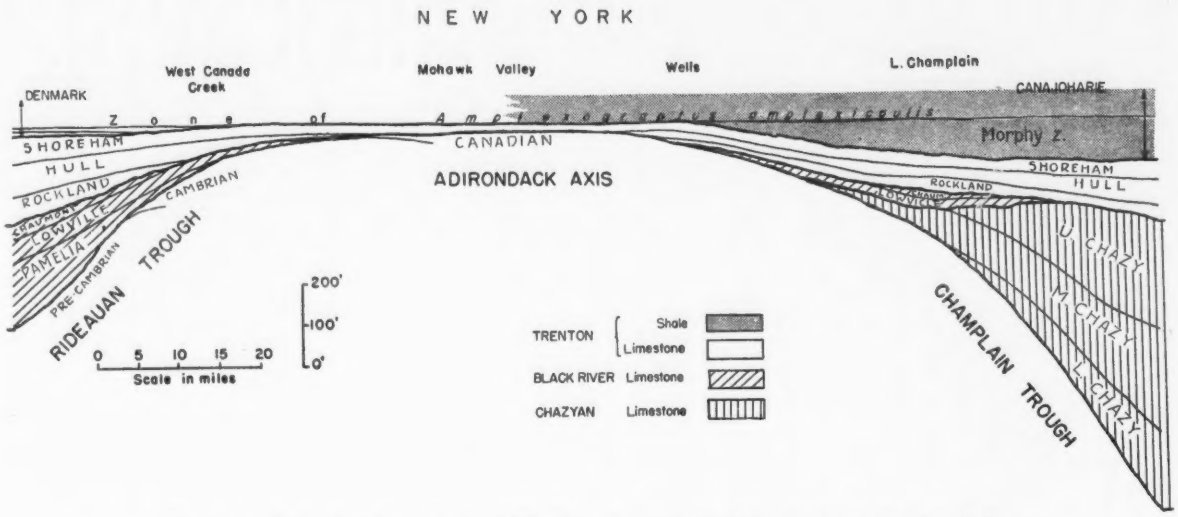


FIGURE 4.—*Restored section of pre-middle Trenton across the Adirondack axis in New York*

The principal representation of the Black River group is west of the Adirondack axis, the older beds being in the Rideauan trough. In contrast, the thick sections of the Chazyan are east of the axis in the Champlain trough, and the series enters the northern Allegheny belt only in the Ottawa Bay. Ulrich (1911, p. 414) considered these relations typical of oscillation; the explanation was somewhat confused by the misclassification of the Pamelia as inter-Chazyan but is fundamentally correct. There is oscillation along a section from Lake Ontario to southern Lake Champlain, in the sense that the Chazyan, thick in the latter area, is absent in the former, and conditions are reversed in the Black River (Fig. 4). There were epeirogenic movements, and the contrast continues southward in the Appalachian region. The Adirondack axis was a hinge, being tilted downward on the east and then on the west.

The lower Trenton Rockland and Hull formations are consistently present in the Allegheny belt and along the west side of Lake Champlain. Rockland seas were at first limited to the Rideauan trough, as evidenced in the distribution of the basal, Selby member (Kay, 1937, p. 252-255; 1942, p. 598-604) (Fig. 5a), and the maximum thickness of the formation, about 70 feet, is northeast of Lake Ontario. If the Amsterdam limestone of the lower Mohawk Valley and southern Lake Champlain be lower Rockland (Kay, 1937, p. 259), there was a similar northeast advance in the Champlain trough. Younger Rockland overlaps the pre-Cambrian north of Lake Huron, Lake Nipissing, Renfrew County, Ontario, and toward Lake St. John, Quebec, where it lies on the crystalline rocks (Raymond, 1916; McGerrigle, 1933). There is similar overlap (Isle la Motte) on older Mohawkian and Chazyan sediments in the western Champlain belt. The distribution suggests that the Champlain and Rideauan troughs were depressed more than the Adirondack belt and that the latter was partially above the sea; the Adirondack arch had come into existence (Fig. 4).

The Hull is consistently present where post-Rockland sediments are preserved; 4 feet represented at Inghams Mills, New York, formerly was not recognized (Kay, 1937, p. 254). The formation has a maximum of about 100 feet in the Allegheny belt and approaches this in the Champlain. The beds overlap Rockland on pre-Cambrian monadnocks and are thought present in the basal Liskeard formation at Lake Timiskaming (Hume, 1925, p. 26). They are more extensive than the Rockland on the Mohawk, but absent on the Adirondack axis at Canajoharie, and are overlapped by the Shoreham near Quebec City on the Laurentide continuant of the axis. Inasmuch as Hull extended beyond the present outcrop, its limits can only be interpreted, but the isopachs reveal an arch on the Adirondack axis with greater depression on the two flanks.

The Shoreham limestone has been described as a member of the Sherman Fall formation (Kay, 1937, p. 264). The beds are mappable as far west as northwestern New York, and inasmuch as the principal stratigraphic contrast within the Trenton group succeeds them, they can be considered better as a third "lower Trenton" formation. As the Shoreham has not been mapped separately in Ontario, use of the term Sherman Fall continues west of New York; regional studies of thickness must be based on the larger unit. The Shoreham is the last Ordovician limestone east of the Adirondack axis. It thins and overlaps older Trenton on the axis and thickens

westward into northwestern New York, reaching about 70 feet in southern Lewis County.

The similarity of Rockland and Hull faunas to those in Arctic America suggests their marine connection (Kay, 1935a, p. 588; Teichert, 1937, p. 44). The Timiskaming Strait (Kay, 1937, p. 289) is not tenable, however, for it is directed toward areas south of James Bay in which Ordovician and Silurian rocks are absent and probably never were deposited, inasmuch as the Silurian and Devonian overlap eastward toward the area (Dyer, 1929; 1930). The strait was conceived in the erroneous belief that the present relief was consequent on pre-Paleozoic elevations (Kay, 1942). If early Trenton seas did pass west of Ungava through western Hudson Bay, sediments of that age have not been recognized. The fauna of the Long Point series of western Newfoundland suggests the presence of lower Trenton (Schuchert and Dunbar, 1934, p. 72); intermigration from the Arctic to the Allegheny belt may have been along the east of Ungava.

The Vermontian disturbance of middle Trenton time raised a highland in the western part of the Magog geosyncline, and this directly influenced the character of the Denmark (Canajoharie) sediments in the Champlain trough. It is reflected in the Allegheny belt only if the marine withdrawal along the northern border resulted from eustatic fall related to the orogeny. That the Quebec axis did not form the place of the Vermontian uplift (Kay, 1937, p. 291) is shown in that west-marginal Magog trough deposits are still preserved in the boulder-bearing Levis shales along the St. Lawrence (Clark, 1923) and the Rysedorph conglomerate at Albany (Ruedemann, 1930), each in the Taconic allochthone; and early Denmark (Snake Hill) shales are present in both the allochthone and autochthone in the latter region (Ruedemann, 1930, p. 29, 117; Kay, 1937, p. 272). The Green-Sutton Mountain anticlinorium, within the Taconic allochthone, has been considered (Kay, 1940) the north limb of the structure responsible for the Vermontian clastics, the land Vermontia. The description of the position of Vermontia is complicated because later orogeny moved the belt in which it formed; Vermontia lay within the western part of the Magog belt, whose rocks have been moved westward so that the allochthonous position of Vermontia approximates the present position of the Green-Sutton axis and of the autochthonous Quebec axis. Later deformation is responsible for some structure of the present anticlinorium, however, there being infolded Silurian sediments along its eastern flank (Clark, 1934).

The disturbance had profound influence on sedimentation in the Champlain trough, which sank nearly a mile beneath the terrigenous sediments now preserved in the Hudson Valley. These shales and sandy shales are limited to the area on and east of the axis (Pl. 1). Relative depression along the sides of the axis is shown in the transgression of older formations by the basal Denmark (Fig. 4), beds older than the zone of *Amplexograptus* being limited to the two flanks, and in the presence of conglomerate-bearing Shoreham pebbles at the base of younger Denmark shales northeast of the Adirondack Mountains (Clark and McGerrigle, 1936).

The Sherman Fall (Shoreham plus Denmark) limestones are about 200 feet thick north and east of Lake Ontario but thin northward through the absence of younger beds. The formation is about 50 feet thick on the Escanaba River, Michigan, Manitoulin Island, and along the Ottawa, and about 100 feet northeast of Montreal; presumably only the Shoreham is represented in the thinner sections. The Cobourg limestone overlaps the Denmark north of a line trending eastward through southern Georgian Bay toward the northern boundary of New York; but the trend becomes northeastward along the Rideauan belt, for there seem

to be thicker Sherman Fall beds north of the St. Lawrence than on the upper Ottawa (Parks, 1929). The reported 100 feet of Trenton at Lake St. John contains lower Trenton and Cobourg elements (McGerrigle, 1933); if present at all, the Sherman Fall must be quite thin.

The depth of water along the north shore of Lake Ontario can be judged, as the elevation of the beds is nearly 100 feet higher on the pre-Cambrian inliers than in the surrounding lowlands (Kay, 1942, p. 632). That the sea was shallower in the Allegheny belt than across the Adirondack axis is shown by pararipples common in Denmark calcite sandstones southeastward to the head of the Black River Valley but not present at Trenton Falls, where coarse interbeds are rare. Between this locality and Canajoharie, the Denmark grades through dense, lime-mudstones into thicker black shales of the Canajoharie formation. The Champlain trough was depressed deeply, permitting the settling of muds that would have remained in suspension had they reached the shallower waters on the west. The belt of gradation is west of the earlier structural arch, and the thickness of deposits and depth of water were greater across the axis than to the west; the Adirondack belt was one of flexure. The thickest clastics are in the Hudson Valley and southwestward into eastern Pennsylvania; the sediments are more calcareous along the Adirondack axis in northern New York than southwest of the mountains. The Vermontian highlands were most rapidly eroded and highest east of southeastern New York.

The principal contrasts in the Cobourg are the northward overlap of the limestone on the beveled Denmark in the Allegheny belt and the limited deposition of equivalent (lower and middle) Utica in the Champlain belt; the eastward continuity of the Utica black shale has not been demonstrated, and, though the Schenectady sandstone of the lower Mohawk has been correlated entirely with the Denmark, and Utica equivalents thought to be absent, the Utica possibly grades into the coarser facies southeastward. The limestones are about 150 feet thick, maintain this thickness northward in the Ottawa Valley, but diminish by offlap toward the Adirondack axis. The lower Cobourg limestone is fully developed as far southeast as near Poland, New York, with coarse calcite sandstones in the upper part; whereas the section above the lower Denmark at Middleville, 8 miles southeast, is entirely black shale, probably including beds equivalent to the Cobourg. There seems to have been a flexure or arch with its axis some 30 miles northwest of the axis of the lower Trenton Adirondack arch. The Cobourg extended across the Canadian Shield if it is represented in the Red River formation of Manitoba and its Arctic equivalents (Kay, 1935a, p. 586; Teichert, 1937, p. 43-44).

In latest Trenton (Collingwood-Gloucester), muddy sediments spread into the Allegheny belt and westward into the region of the present Michigan basin. This marked change in lithology has been interpreted as resulting from the filling of the Champlain trough, so that the terrigenous materials from Vermontia were carried into deepened waters transgressing the buried Adirondack arch. The relations initiated continued into the Cincinnatian epoch.

Subsurface Mohawkian.—In the Allegheny synclinorium and westward, the Black River and pre-Collingwood Trenton formations are limestones, contrasting with the argillaceous overlying beds and with the underlying crystallines and marginal sandstones. This lithologic unit is separated by the driller as the "Trenton lime." An isopachal map of these rocks must be based on a limited number of wells that have penetrated them, many wells being concentrated in small areas that yield oil and gas. The

map (Fig. 5b) is limited eastward by the Adirondack belt in which the limestones grade into shales in the middle Trenton and are partly cut out by unconformity in the Cobourg.

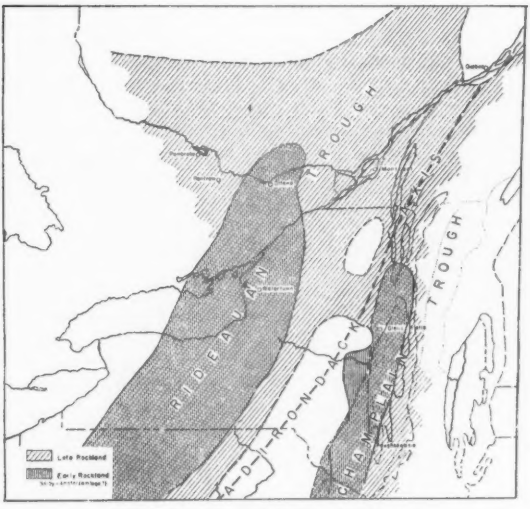
The first factor affecting thickness is the relief on the pre-Paleozoic surface. As a whole, this must have been quite small, for few wells give thicknesses varying widely from the average of many. Thus, of 15 wells in Kent and Essex counties, Ontario (Harkness, 1920-1936; Knight, 1915), the minimum is 790 feet, the maximum, 948; but only four vary more than 25 feet from the average of the whole number. The relief is well known on the outcrop, where partially exhumed monadnocks are scattered (Kay, 1942, p. 629-632).

A relatively thick belt extends from the Ottawa through Madison County, New York (Fig. 5b). The outcrop belts traversed are in the Rideauan trough in which the Pamelia is thickest; this factor accounts for most of the contrast with adjacent areas. A second trend of maximum thickness passes northwestward through Kent and Lambton counties, Ontario, lying along the Chatham sag in the Cincinnati arch, and evidences an early origin of this cross structure; this may be a belt in which the Black River is also thicker and more complete.

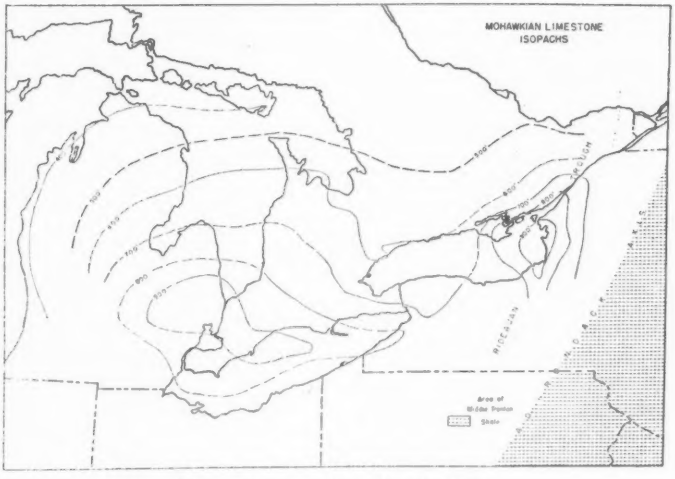
The thickness diminishes northward, due partially to overlap of the Black River group. Additional cause is the Sherman Fall-Cobourg disconformity, which results in the loss of some 150 feet of limestone between Lake Ontario and a line eastward from northern Lake Huron. There is some convergence in other formations, but it is not of great significance.

Thus the varying thickness of the Mohawkian "Trenton lime" reflects (a) irregularities on the pre-Ordovician surface, (b) the deposition of earlier rocks restricted to the Rideauan trough and possibly the Chatham trough with marginal overlap, and (c) the overlap of the Cobourg on diminishing Sherman Fall.

Cincinnatian series.—The sediments of the Cincinnatian in the Allegheny synclinorium are argillaceous and arenaceous, in contrast to the carbonates of the Mohawkian (Pl. 1). They are best exposed east of Lake Ontario (Ruedemann, 1925), are concealed westward beneath the lake, and sparsely shown along the outcrop north and east of the Niagara escarpment in New York and Ontario (Dyer, 1925; Parks, 1925; Fritz, 1926). Much of the understanding must be gained from logs of wells that have penetrated the series. The Cincinnatian converges northwestward (Fig. 6a), being 1000 feet along an arcuate line trending from the north end of Lake Simcoe to St. Thomas, Ontario, and from Sandusky toward Bellefontaine and Portsmouth, Ohio. The 1500-foot isopach lies about 50 miles eastward, near Hamilton and Port Dover, Ontario, Cleveland, Ohio, and in about western Roane County, West Virginia. The thickness exceeds 2000 feet in south-central New York and central Pennsylvania. The rocks form a low semi-cone of terrigenous sediments, having an apex in eastern Pennsylvania and evidencing a semilenticular depression of their floor.



a. Distribution of lowest Trenton (Rockland).



b. Isopachal map of Mohawkian limestones.

FIGURE 5.—*Mohawkian paleogeography*

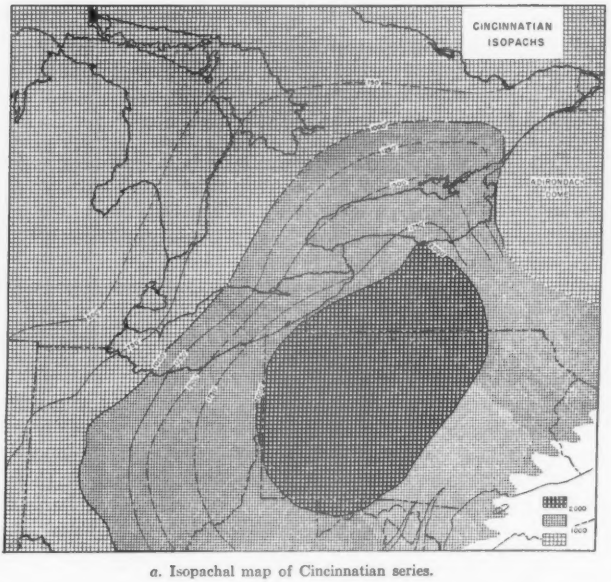
The series has two principal facies in the northern part of the synclinorium—the lower gray shales and sandy shales, and the upper red shales (Queenston) of Ontario and Ohio that pass into coarser red beds (Juniata) southeastward. The red facies gradationally overlaps late Cincinnatian (Richmond) gray shales northwestward and is overlapped by Silurian conglomerates southwest of the Adirondack Mountains and in northeastern Pennsylvania. Although the sub-Richmond Lorraine or Eden-Maysville converges northwestward, the addition of lower Richmond shales of similar lithology gives the facies a fairly constant thickness of about 1000 feet in New York, Ontario, and northern Ohio. The Lorraine retains this thickness to the limit of its outcrop west of the Adirondack Mountains (Kay, 1939); it probably continued over the Adirondack axis without significant change, possibly with expansion, and across the Champlain trough to the source in Vermontia.

There was late Maysville uplift southeast of central and eastern Pennsylvania, for the sandstone and conglomerate of the former area exceed 1300 feet. Thin conglomerate at the base of the Shawangunk (Silurian) conglomerate that has been correlated lithologically with the Bald Eagle (Maysville) conglomerate lies unconformably on the older Lorraine in eastern Pennsylvania (Willard and Cleaves, 1939); the correlation seems tenuous. The effect of the disturbance is reflected in the coarse texture of the uppermost Lorraine, Oswego sandstone, which spread as far as central New York and northern Virginia. The disturbance did not extend far northward as shown by the absence of sandy facies in the thick section on the Nicolet River south of Three Rivers, Quebec (Foerste, 1916, p. 14-41). Moreover, the great thickness of the Lorraine in that area shows continued depression of the northern Champlain trough.

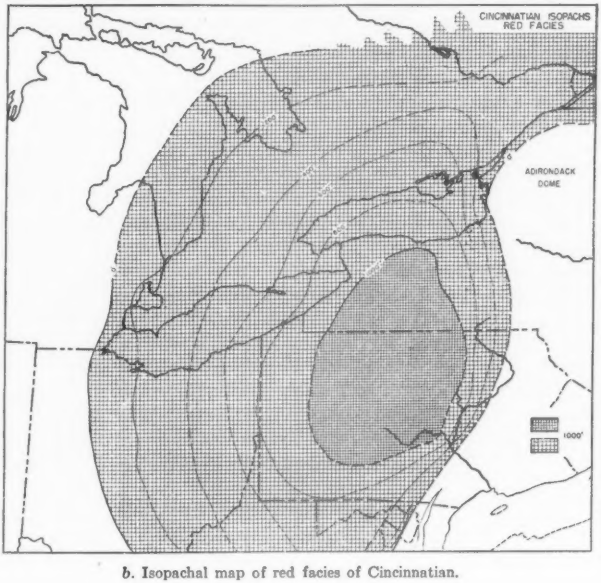
The Queenston red facies (Fig. 6b), the base rising stratigraphically northwestward by gradational overlap, extends to eastern Ontario, Lake Huron, southeastern Michigan, central Ohio, and southern Virginia with an arcuate margin; gray shales with marine faunas comprise the whole of the Cincinnatian peripherally. The Richmond shales overlap the Lorraine between Manitoulin Island, where both Eden and Maysville are present (Caley, 1936), and northern Michigan (Hussey, 1926). The Richmond (Juniata) red beds are absent in eastern Pennsylvania (Willard and Cleaves, 1939), extending only into the western margin of the Champlain belt, and similar relations pertain southwestward into Virginia (Butts, 1933). The equivalent Queenston is overlapped by Silurian southwest of the Adirondack Mountains.

The area of the Adirondacks and of eastern Pennsylvania lay to the east of the graded plains of the sinking basin in the Allegheny Belt. In southeastern Quebec and westward through Ontario, marine waters persisted through early Richmond (Waynesville); and more than 1000 feet of fine-textured Queenston in Quebec was laid in the sinking Champlain trough. The seas retreated before the advancing terrigenous sediments, producing the gradational overlap of the red shales over the gray marine shales. Isopachs of the red facies (Fig. 6b) do not represent synchronous deposits but support the view that the principal depression was toward eastern Pennsylvania.

The Cincinnatian isopachs cross those of the Trenton in western Ontario. There is no appreciable abnormal thinning of the sediments on



a. Isopachal map of Cincinnatian series.



b. Isopachal map of red facies of Cincinnatian.

FIGURE 6.—*Cincinnatian isopachal maps*

the east flank of the Cincinnati axis, and the facies lines do not evidence influence of an arch or flexure. The Cincinnati arch is believed of later origin.

TACONIAN REVOLUTION

The Taconian revolution [2] at the close of the period involved thrusting of rocks deposited in the Magog geosyncline on the folded strata of

FIGURE 7.—*Structure section from Lake Ontario to New England*

the Champlain trough; the overthrust rocks form the Taconic allochthone (Figs. 1, 7). The principal area of allochthone is the Taconic klippe, composed of argillaceous Cambrian to middle Ordovician lying as a tectonic outlier of stratigraphically exotic rocks surrounded by equivalent carbonates of more westerly origin in eastern New York, southwestern Vermont, and western Massachusetts. Similar "outliers" have been demonstrated allochthonous in western Newfoundland (Johnson, 1940); the anomalous character of the facies and faunas suggests this relationship, though the beds were considered older than the carbonates previously (Schuchert and Dunbar, 1934, p. 38-40). The Taconic thrust evidently lies along White Bay, for the sediments on its west (Betz, 1939) are of Champlain trough facies, those to the east in Notre Dame Bay (Heyl, 1937) of the Magog. In Pennsylvania, similar exotic shales have been interpreted as possibly in a klippe of the Martic thrust sheet (Kay, 1941a). The minimum displacement, disregarding the effects of retreat of its most advanced front through erosion, foreshortening in later movements (Cloos, 1940), and the gap that must have separated the westernmost preserved allochthone and easternmost exposed autochthone, was about 40 miles in Newfoundland and in New York-Massachusetts (Kay, 1937, pl. 5; Schuchert, 1937, p. 1028), and 30 miles in Pennsylvania. The true displacement was considerably more.

Inasmuch as the marginal allochthonous sediments are of Magog belt facies, the fault was initiated within the border of that geosyncline. That it was adjacent to the Quebec axis is shown by conglomerates with exotic

[2] The writer and others have referred to the post-Ordovician orogeny as the Taconic disturbance (Kay, 1937, p. 291; 1940), a term less desirable than Taconian revolution of other authors; the use of Taconic for both a thrust sheet and an orogeny leads to confusion, and the lateral movement, one of the greatest in North American history, warrants the term revolution.

(Champlain belt) boulders in Quebec (Clark, 1924) and New York (Ruedemann, 1901; 1930). In Maryland, the Champlain belt sediments near the Martic thrust thin and diminish in content, as though by overlap (Stose and Jonas, 1936).

The thrust sediments are in tectonic contact on Queenston shale in southeastern Quebec, and the autochthonous Cincinnatian has been folded considerably (Parks, 1931, p. 29-30). The overthrust rocks are overlain at Becraft Mountain, New York (Kay, 1937, p. 287-288) and in the Catskill Front (Ruedemann, 1930, map) by latest Silurian Manlius limestone. Thus, there is direct evidence that the principal lateral movements were pre-Manlius and post-Queenston. Folds in autochthonous Ordovician are truncated by the Shawangunk and Tuscarora quartzites of the earliest Silurian in southeastern New York and Pennsylvania; if the folding accompanied Taconic thrusting, the revolution is pre-Silurian.

The front of the thrust sheet was not very high. Middle Ordovician sediments are preserved near to the westernmost remnant of the sheet and probably never were buried deeply. On Anticosti Island in the Gulf of St. Lawrence, there is essentially continuous section of Cincinnatian and early Silurian calcareous shale and limestone (Twenhofel, 1927) in the Champlain belt within 50 miles of the overthrust rocks of Gaspe; the allochthone was beneath the sea or not high enough to produce significant detritus after the revolution. Though the quantity of Silurian terrigenous sediments is distinctly smaller than that of the Ordovician, as will be shown later, this reflects repeated uplift and continued presence of Vermontian highlands in later Ordovician, in contrast to progressive reduction of the transposed Taconia in the Silurian. The greatest quantity of eroded material was laid in the latitude of Pennsylvania, as shown by isopachs; that the greatest elevation was there also is shown by the coarser texture of the sediments. The lateral movement of the allochthone may have been as great or even greater in Quebec, but Vermontia and its transposed descendant, Taconica, were more continually high farther south. Quantity of erosion is measured by volume of sediments; elevation and proximity of source by texture.

The metamorphic pre-Cambrian rocks on the sole of the thrust where it is exposed farthest east in Pennsylvania probably show preceding uplift in the core of the Vermontian geanticline. It has been suggested that the thrust passes west of the Green Mountain anticlinorium east of Rutland, Vermont (Hawkes, 1941); the metamorphic rocks may correspond to the Glenarm sediments of Manhattan and southwestward. The presence of later Ordovician slates from Pennsylvania to Virginia in the Martic thrust block implies that the axis of Vermontia lay west

or east of the metamorphics adjoining the fault, and their structure evidences post-Vermontian compression. The hundreds of cubic miles of detritus that came from Vermontia and Taconica demand raising of a considerable belt within the Magog geosyncline; the metamorphics would represent what lay beneath the floor of the Paleozoic geosyncline. If Ordovician rocks are to be found, they should be farther southeast and correspond to those of the New Hampshire synclinorium (Billings, 1936) whose western limb is within the Magog belt.

"Appalachia" is a term applied to undefined land southeast of the Appalachian region sediments, just as "Llanoria" is utilized in south-central United States (Van der Gracht, 1931a, p. 118) and "Cascadia" in the west. Referring to the rocks of the Martic anticlinorium—"the embryonic ranges which formed as a result of the first post-Cambrian orogenic deformation arose in the eastern half of the Ordovician geosyncline" (Bucher, 1933, p. 132)—Vermontia developed in the Magog geosyncline, within the Appalachian belts. The Glenarm was laid in a pre-Cambrian geosyncline, evidently the precursor of the Magog, but it was not deformation at the close of the Cambrian (Jonas, 1929, p. 510) that elevated the rocks, but the Vermontian disturbance in the Ordovician; one must not infer that the Martic anticlinorium is the Vermontian geanticline, for it was later deformed.

SILURIAN SYSTEM

Niagaran series.—The earlier Silurian Niagaran series is separable into the Alexandrian, Clinton, and Lockport groups, each present in the synclinorium (Pl. 1). The inclusion of all pre-Cayugan Silurian in a single series (Grabau, 1921, p. 314) seems advantageous; the typical rocks formed with essentially progressive overlap, though incomplete at the base and within in New York State. Prior to the Taconian revolution, the region of the present Adirondacks formed a broad dome in late Cincinnatian. Following the orogeny faulting along the east flank produced the present block-mountain structure. The time of faulting has not been conclusively established, but is considered early Silurian (Megathlin, 1939, p. 119-120); what seemed most significant, the considerable displacement on the Mohawk and its seeming absence in the Silurian of the Helderberg Plateau near by (Kay, 1937, p. 288) has been found due to termination of the faults rather than their being overlapped (Megathlin, 1939). On the northwest side of the synclinorium, the Alexandrian and particularly the Clinton evidence establishment of an arch along the Cincinnati axis. Thus the western limit of the Allegheny belt had become established in early Silurian.

The Albion or higher Alexandrian has been studied thoroughly along the Niagara escarpment in Ontario (Williams, 1915a; Cumings, 1939) and in wells south and southwest of the outcrop (Harkness, 1928; Newland and Hartnagel, 1932, 1936; Fettke, 1935; Hartnagel, 1939). The beds change from dominantly sandstone in north-central New York to limestone and intercalated shale in the Michigan basin (Pl. 1). The sandstones are limited to the southeast of the present Cincinnati arch, practically disappearing northwest of a line from Port Stanley through Brantford toward Toronto (Harkness, 1929, p. 67-68); they are similarly restricted to the east flank of the arch in Ohio (Ballard, 1938, p. 1534). Seas passed over the axis and are represented principally by carbonates in the Michigan basin. North of the extension of the axis, at Lake Timiskaming, beds have facies and faunal characters of the Albion of the Michigan basin (Hume, 1925, p. 28-33).

Eastward in New York, the Albion is overlapped by the Clinton, the Oneida conglomerate (Sanford, 1936, p. 803-804) forming the marginal facies. The sandstones retain thickness of about 100 feet through western New York, approaching 200 feet in northern Pennsylvania wells. In central Pennsylvania, the Tuscarora quartzite is nearly 1000 feet; only the upper, Castanea, member is thought equivalent to the New York Albion (Swartz, 1934, p. 111). The older Alexandrian sediments must converge and overlap rapidly in north-central Pennsylvania. Eastward in Pennsylvania and in southeastern New York, the Alexandrian is contained in the lower part of the thick, conglomeratic Shawangunk formation, separated from the Albion of central New York by the south-plunging Adirondack arch. The arch shielded the northern Allegheny belt from direct influx of sediments from the highlands of Taconica. Rapid erosion is implied by the coarse texture and quartz content of the detritus. The trough in the Allegheny belt plunged gently southward in southern New York, the floor deepening rapidly below the early Silurian semilenticular mass in eastern Pennsylvania.

Clinton sedimentation simulated that of the Albion, but the group is absent along the Cincinnati arch in Ontario east of Lake Huron, and a contrast in facies is represented on the flanks. The Clinton disappears across the "Cataract Axis" (Fig. 4) (Schuchert, 1911; Williams, 1915b) along the trend of the Cincinnati arch, the structure evidently plunging southwest toward the Chatham sag, where the Clinton breaches the axis. The margin of Clinton shales runs from St. Marys through Stratford and Kitchener (Harkness, 1929, p. 66). Northwest of the peninsular arch, in the Michigan basin limestones extend southeastward to the south end of Georgian Bay (Cumings, 1939, p. 596). Though Clinton is not reported at Lake Timiskaming, beds may have been included in the lower part of the "Lockport", as has been done in the past in Michigan.

Thus, the Cincinnati arch is first clearly indicated in the Clinton. This advent is about synchronous with that in Ohio, where Albion sediments persist but are thin on the arch, whereas the Clinton offlaps (Foerste, 1935). That the arch was not present in the Ordovician has been indicated from the outcrop and isopachs; the Chatham sag was suggested by Mohawkian isopachs, however. The evidence for post-Ordovician age in Ohio and Kentucky has been summarized by McFarlan (1939); structural trends during pre-Mohawkian and Mohawkian time are across the trend (McFarlan, 1938). Moreover, the Cincinnatian is

quite fully developed on the axis where the lower Silurian is poor and incomplete.

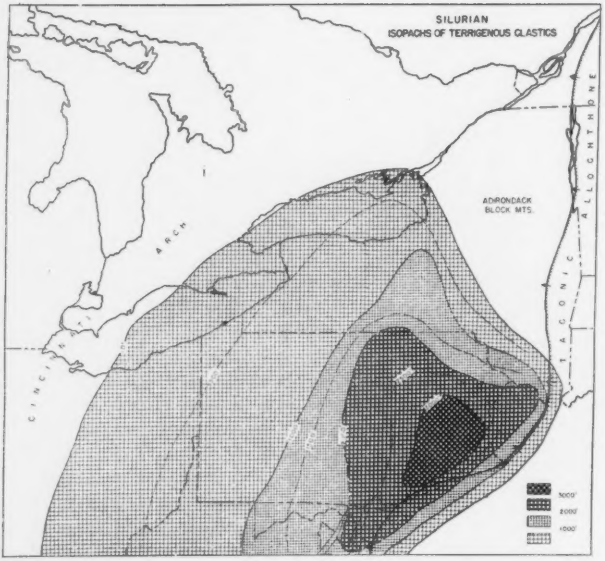
Clinton sediments in New York have lensing section, thicken eastward to Onondaga County, and change to coarser facies eastward, the latter, the Oneida, disappearing in Herkimer County (Chadwick, 1915; Sanford, 1936). There is northward convergence from 800 feet in central Pennsylvania (Swartz, 1936) to 200 feet in western New York and about 300 feet in Onondaga County. The character of the Oneida, with white quartz pebbles, implies an origin similar to the partially equivalent Shawangunk; the pebbles evidently came around the south flank of the Adirondack arch.

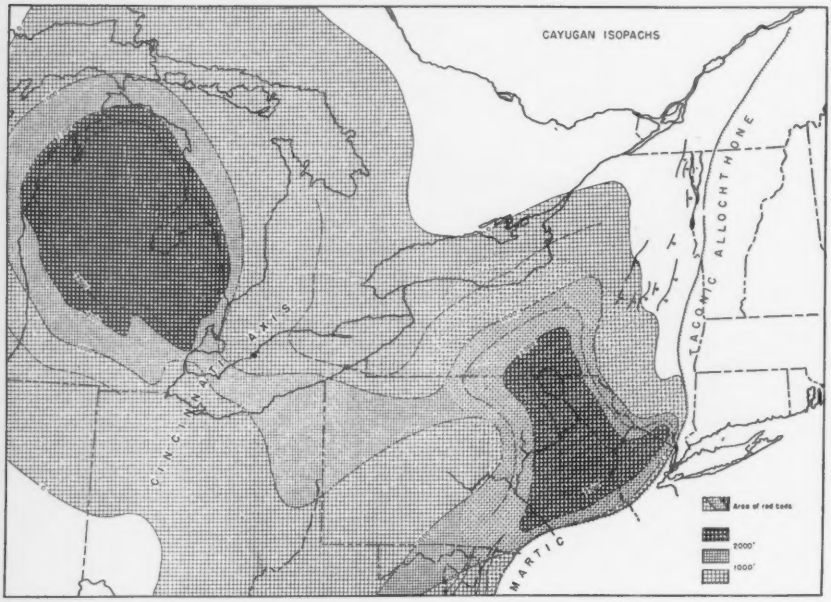
The Lockport group is persistently carbonate and rather constantly 150 to 250 feet thick, from Georgian Bay to western New York, thinning and disappearing by offlap in Onondaga County. The lower Lockport (Manistique) is overlapped southeastward from the Michigan basin (Cumings, 1939), the basal dolomite in southern Ontario and New York being the Engadine, lying disconformably on Clinton. The beds become argillaceous south of the deepest wells in central New York if the McKenzie shaly limestone of central Pennsylvania corresponds (Swartz, 1934), and somewhat thicker; it has not been determined whether these include pre-Engadine Lockport and are basally overlapped northward. The McKenzie passes eastward into the lowest red shales of the Bloomsburg of eastern Pennsylvania (Swartz and Swartz, 1931).

The Adirondack belt in New York lacks deposits older than latest Silurian; evidently it was low, for the Lockport is not terrigenous to the west. Lockport blocks have not been found in the volcanic breccia at Montreal; possibly there was an arch as to the south of the Adirondack Mountains.

Cayugan series.—Cayugan rocks are exposed in three isolated areas: along the southeast of the Michigan Basin; north of eastern Lake Erie and south of Lake Ontario along the northern margin of the Allegheny synclinorium; and in the folded Appalachians of Pennsylvania, Maryland, and the Virginias. Exact correlations are uncertain, but the gross deformation can be ascertained, for wells penetrate the series in southern New York, eastern Ohio, and southwestern Ontario (Pl. 2b). The highest Cayugan Manlius limestone of New York seems equivalent to the Keyser limestone of the Appalachians (Swartz, 1931; 1939); these beds are absent in western New York and southern Ontario but are thought equivalent to the upper Bass Island group in western Ontario (Williams, 1915, p. 93-94).

From Erie and northern Chautauqua counties, western New York, to Elgin and southern Oxford counties, Ontario, the series is composed of 350 to 400 feet of gypsum-bearing dolomites (Fig. 8). Westward, the rocks thicken to more than 1000 feet near Detroit, and exceed 2000 feet in Livingston County, Michigan. Part of the thickening is by addition of higher beds, but it is considerably due to divergence, an indication of the sinking character of the Michigan basin region





SILURIAN ISOPACHAL MAPS

a. Isopachal map of terrigenous sediments in the Silurian system

b. Isopachal map of Cayugan series.

and the adjoining Cincinnati axis. Eastward in New York, it thickens to more than 1000 feet in Onondaga County, then thins, and the Manlius overlaps on pre-Silurian beds east of Herkimer County over the Adirondack axis. Along this section, the basal Cayugan Vernon red shale enters in Genesee County, exceeds 500 feet in Onondaga County, then disappears eastward; it is succeeded by the principal salt beds.

The isopachal map (Pl. 2b) shows a southwestward-plunging basin in New York west of the Adirondack axis, the Cayugan thickening to more than 2000 feet in eastern Schuyler County, where the lower 800 feet is of Vernon red and green shale (Hartnagel, 1939). More than 2800 feet of Cayugan has been penetrated in Tioga County, northern Pennsylvania, but the total decreases westward to less than 1000 feet in southern Cattaraugas and Chatauqua counties, southwestern New York. The maxima then trend southeastward into eastern Pennsylvania, where nearly the whole section is Bloomsburg red shale, exceeding 2000 feet, with succeeding upper Tonoloway and Manlius limestones (Swartz and Swartz, 1931); in central Pennsylvania there is about 1200 feet of Cayugan (Butts and Moore, 1738; Butts, 1939).

There is no information on subsurface Cayugan through western Pennsylvania. Across West Virginia, the Albion to Oriskany interval, including beds additional to Cayugan, decreases from 2400 feet in Randolph County to less than 1600 feet in Roane, and 1435 in Wood County (Martens, 1939), suggesting convergence of the Cayugan that comprises most of the unit. Increase to 1700 feet in Harrison County, Ohio (Ballard, 1938), suggests that the Allegheny belt was a broad arch separating the Champlain belt from the Michigan basin as in the north.

The tongues of lower Cayugan red shales extend southwestward into northern Virginia in the base of the Wills Creek, and in New York to Genesee County; their limit is arcuate. Saline rocks lie northwest of the lower Cayugan red shale semicone; southward the limy shales of the Wills Creek are marine. Flow of marine water to the interior was restricted by the Bloomsburg-Vernon red shale delta and the relatively shallowly flooded Allegheny belt in West Virginia, Pennsylvania, and southern Ontario, isolating the northwestern Allegheny belt and Michigan; precipitates formed in these partially enclosed, sinking interior areas. The two saline basins were partially separated along the middle of the Allegheny belt. The basin extended northwestward to the west of Ungava, for there are dolomitic shales southwest of Hudson Bay (Dyer, 1930).

The youngest Cayugan limestones overlap the Adirondack axis in eastern New York and extend eastward over the margin of the Taconic allochthone. They are present in the eastern border of the Allegheny belt, and in the Champlain belt, and may be represented in the upper Bass Island carbonates of the Michigan basin. The Allegheny belt was less depressed than the adjoining geosynclines.

There is much less terrigenous material in the post-Taconian Silurian "molasse" than was derived from Vermontia in the later Mohawkian

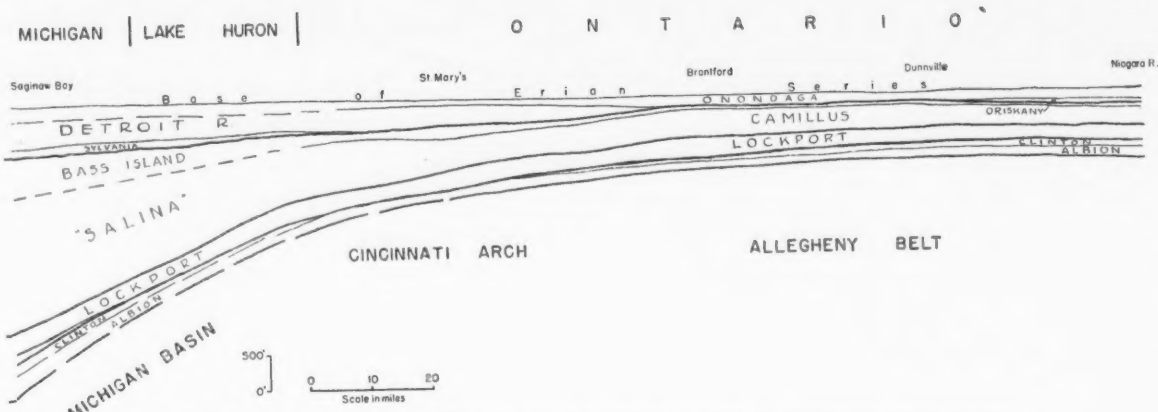


FIGURE 8.—*Restored section of Silurian and older Devonian across the Cincinnati axis in Ontario*

and Cincinnatian (Fig. 6a; Pl. 2a). Within the area of preservation east of the Cincinnati axis in Ohio, Pennsylvania, Ontario, and New York, the quantity of the former approximates 65,000 cubic miles, of the latter, 95,000. The Ordovician quantity should be almost doubled, for it excludes a great volume of Mohawkian and Lorraine in the Champlain belt and a smaller amount of Cincinnatian shale west of the Cincinnati axis.

DEVONIAN SYSTEM

Helderbergian, Oriskanian, and Ulsterian series.—The Lower Devonian Helderbergian and Oriskanian series are restricted to the area east of the Cincinnati axis. The latter are the oldest rocks penetrated by many wells within the synclinorium in Pennsylvania. The greatest depression was in a trough having more northeasterly trend than the Ordovician Adirondack axis, and the margin of the sea was within the western Champlain belt; the structural influence that directed the form and trend of the Adirondack axis had been superceded. The Oriskanian evidences the relatively higher structure of the Cincinnati belt and the separation of the Allegheny basin from the St. Lawrence, possibly along the Frontenac axis.

Helderbergian limestones were laid disconformally on the Manlius in eastern New York and Pennsylvania. They continue across the Adirondack axis into the Champlain belt with eastward thickening by offlap in New York and are absent in central and western New York. They are nonterrigenous east of the Hudson at Becraft Mountain and represented by limestone blocks in the St. Helens volcanic breccia at Montreal, signifying that the Adirondack area and Taconic allochthone were rather low. The Allegheny belt was tilted southeastward; the axis of the New Scotland (Swartz, 1939, p. 57) and earlier troughs cross the Adirondack axis in Pennsylvania. The limestones pass into marginal sandstones in western central Virginia (Swartz, 1929), within the Champlain belt.

The margin of the Oriskanian is rather clearly delimited in southern New York and northwestern Pennsylvania (Fig. 9) (Fettke, 1938, p. 241; Lafferty, 1938, p. 175; Martens, 1939, p. 30-35) and along the northwest; it paralleled the Cincinnati axis, approaching most closely at the Chatham sag. The sands were laid in a shallow, southward-deepening sea and are locally absent within the synclinorium, as in southern McKean County, Pennsylvania (Fettke, 1935, Fig. 1). They extend eastward into the Champlain belt in New York and Pennsylvania (Cleaves, 1939). Oriskanian blocks in the St. Helens breccia at Montreal (Williams, 1910) indicate that the Adirondack region had become connected to the shield; the margin of the sea that trended eastward south of Lake Ontario evidently turned northward around the east flank of the Adirondacks and was embayed in the upper St. Lawrence basin. The trend and character of the Devonian arch cannot be delimited but may have been in the position of the Frontenac axis.

The Ulsterian is more extensive, the Onondaga limestone in the Allegheny synclinorium and the Detroit River series of the Michigan basin being representatives (Warthin, 1939, p. 601-602). It forms the last dominantly carbonate sediment in

the region, the advent of Acadian orogenic movements in the Erian leading to the formation of westward-thinning terrigenous sediments of the later Devonian (Cooper, 1931; Reeves and Davies, 1937). The Onondaga persists with a thickness of 100 to

FIGURE 9.—*Margin of Oriskanian sediment in the Allegheny belt*

200 feet westward to the Cincinnati axis; thence Ulsterian carbonates increase to nearly 1000 feet in the Michigan basin, perhaps partially through overlap (Newcombe, 1933). Limestones continue southward into central Pennsylvania from New York without significant thickening; northeastwardly along the outcrop, they diminish across the "Harrisburg" axis or peninsula in east-central Pennsylvania (Willard, 1939) and exceed 500 feet in northeastern Pennsylvania, the lower beds being shale. Thus, in the Allegheny belt, the Ulsterian forms a plate of limestone, increasing in the western Champlain belt with addition of basal shales and diverging greatly into the Michigan basin.

Later Devonian series.—The post-Ulsterian sediments in the synclinorium are almost wholly terrigenous. The stratigraphy of the Erian in New York (Cooper, 1930) and succeeding Tully (Cooper and Williams, 1935) and Genesee (Grossman, 1941) have been described, as has that of the equivalent beds in Pennsylvania (Willard, 1939). There is general gradational overlap of the more sandy facies on the more shaly northwestward, with accompanying convergence (Reeves and Davies, 1937). The culmination is the spread of the continental facies in the late Devonian (Barrell, 1913-1914; Caster, 1934; Chadwick, 1933; 1936; Willard, 1939). Isopachal maps for Pennsylvania (Ashley, 1938) and West Virginia (Tucker, 1936, Pl. 1) emphasize the southeasterly tilting of the northern Allegheny synclinorium as the northwest quadrant of a great semicone of terrigenous materials. The source was highest and most eroded east of northern Pennsylvania.

SYNTHESIS OF EARLIER PALEOZOIC DEFORMATION

STRUCTURAL BELTS

The "major linear folds" of the present structure have been recognized and discussed by Keith (1923, p. 325); these have northeast trends as far west as the Cincinnati arch. In Ordovician and Silurian time, similar structural trend was dominant, but the alternate belts of relatively higher and lower structure have been considerably obscured by later deformation and burial. On the west, the Cincinnati axis, first structurally evident in the Silurian, was an arch only for a limited time and is on the margin of the Michigan basin. The Allegheny belt was essentially a structural plateau, the Rideauan trough within its eastern border being temporarily depressed. The Adirondack axis, first shown in the middle Ordovician, had Ordovician structure similar to that of the later Cincinnati axis; for part of Mohawkian time it was the site of an arch, but at other times the foreland margin of the deepening trough to the east. In late Ordovician (Richmond), it became domed relative to the surroundings in the region of the present Adirondack Mountains; normal faulting along the east flank and of the Laurentide dome to the northeast formed the present block-mountain structure, probably in early Silurian. Southwestward from New York, the Ordovician structural character has been obscured by later deformation and is not evidenced in younger sediments.

The Champlain belt to the east was a geosyncline until late Ordovician Cincinnatian and continued to subside more than its foreland. It was mostly destroyed and obscured by the Taconian orogeny, but the belt was overlapped from the northwest by later seas. Thrust sheets have buried the Quebec axis which is interpreted as having separated the

Champlain from the Magog trough to the east. Inasmuch as the Magog belt was thrust on the Champlain, their original relative positions can be represented only on a palinspastic map (Fig. 2). Thus, there were from west to east the following northeast-trending structural provinces or paleogeographic belts: Cincinnati, Allegheny, Adirondack, Champlain, Quebec, and Magog, differing in character and time of initiation.

GROSS DEFORMATION

The total deformation was downward in all parts of the Allegheny, and contiguous belts through early and middle Paleozoic time, probably excepting the higher parts of the Silurian-formed Adirondack and Laurentide block mountains; that is, with minor exceptions, the surface on which Ordovician sediments were laid was far below the sea level of late Devonian time. The structurally higher areas only moved down less than the structurally lower. The principal sources of clastic sediments consistently lay to the southeast, being east of the Magog trough until at least late Chazyan time in the central Appalachians, until middle Trenton in the northeast. They were in the western Magog belt in the later Ordovician, within the thrust Taconian block in about the position of the autochthonous Quebec axis in Silurian, and in similar position but newly formed elevations in the later Devonian. The pre-Ordovician surface east of the Adirondack axis had been considerably deformed by late Devonian; much of that area, if not all, was above Devonian sea level.

REGIONAL TILTING

The Allegheny belt was tilted as a whole southeastward during times when uplift farther east mantled it with detrital sediment. Isopachs for each of the great clastic series are somewhat arcuate, the rocks forming the northern part of semiconical forms that center in the Champlain belt in eastern Pennsylvania. These reflect the presence of highlands to the east but also the influence of the Adirondack dome of the late Ordovician and the subsequent block mountains in inhibiting westward spread of sediment from New England. Depression was synchronous with deposition, and greatest where clastic sediments were thickest; deposition exceeded depression in latest Ordovician and late Devonian, prior to the Taconian and Acadian orogenies, causing shores to retreat northwestward from the advancing fluviatile plains. When terrigenous material was of limited amount, tilting tended to be southward, away from Laurentia, and marginward toward adjoining troughs.

Inasmuch as the greatest part of the thickness in the northern Allegheny synclinorium is in these Cincinnatian, Silurian, and Devonian

detrital sediments, the northern part of the belt was tilted southeastward as the northern part of the semilenticular depressions.

THE AXES

The Adirondack and Cincinnati axes had arches for limited times (Pl. 3). The Adirondack arch is evidenced only during part of the Mohawkian epoch, and by domes in the late Cincinnatian, and the Cincinnati arch during the early Silurian Clinton, and possibly in late early Devonian. The deformation along the northern part of the Adirondack axis, in the Adirondack and Laurentide areas, formed block mountains in the early Silurian, and these stood persistently higher than the seas. The rock surface along the Cincinnati arch never stood markedly higher than on its flanks.

The axes margin an intervening Allegheny belt that tended to be less depressed than the Champlain trough and the Michigan basin except during orogenic times. Sediments tend to be thicker on the axes than in the intervening belt when adjoining troughs were greatly depressed, but the axes of present structures approach those of the early Paleozoic arches. In middle Trenton time, similar thickness of limestone was laid throughout the Allegheny belt along lines normal to it, but the beds on the Adirondack axis are terrigenous and thicker, diverging into the Champlain trough. Similarly, Cayugan carbonates are of fairly uniform thickness across the central part of the belt, but thicker on the Cincinnati axis, diverging into the Michigan basin. The Adirondack axis is one of oscillation in the Ordovician. Chazyan sediments are principally in the Champlain belt, Black River in the Rideauan trough of the Allegheny; the principal depression was on alternate sides. The Champlain belt contains much thicker sections of the middle Trenton; the gross result of these alternations is to have made the total thickness greater on the east flank than on the axis. Inasmuch as the Cincinnati axis separates the sinking Michigan basin from the southeast tilting along the margin of the semilenticular terrigenous masses of the Allegheny belt, there is alternation in maximum depression on the two sides between nonorogenic and orogenic times, resulting in less gross depression on the axis than on either flank.

The disturbances along the Cincinnati axis in Tennessee have been discussed by Wilson (1935). He concluded that, when "Appalachia" was relatively high, the Nashville dome was completely above sea level, for most of the time, or submerged only on the outer flanks, and that when "Appalachia" was low the dome was relatively low and was probably below sea level for most or all of the time. These relations might result from synorogenic upwarping of the dome or from eustatic with-

drawal from an area of relatively higher surface; he attributed the sympathetic synchronism to the former. The sequence for the "Appalachian geosyncline" of Wilson is composite; some formations, as those of the Blount group, are in the Appalachian allochthone and were laid in the Champlain trough; others, as the "Trenton limestone," are nonterrigenous only in the Allegheny belt. McFarlan (1939) has questioned the presence of an arch in Kentucky in pre-Silurian and has shown that earlier structures were transverse (1938).

In Ontario, the first evidence of arching is subsequent to the Taconian revolution, though when highlands remained. Later history reveals some contrasting movements. The lower Lockport (Manistique) is limited to the Michigan basin and overlapped eastward by the Engadine, which forms a "plate." The Cayugan converges eastward across the axis, being thinnest in the middle of the Allegheny belt, evidencing depression of its margins. The late Cayugan probably offlaps each flank of the axis. The Helderbergian and Oriskanian are absent proximal to the axis, but the latter's margin in the Allegheny belt conforms to present structure (Fig. 9). Ulsterian limestone converges toward the axis from the west (Warthin, 1939) and probably overlaps. There being no recognized disturbance between the Taconian and mid-Devonian to the east, the secular movements cannot be called synorogenic, and the axis was more depressed than the belt to the east except in the early Devonian.

The basal Erian (Marcellus) is limy on the west, argillaceous to the east, and, though the contrast is not marked, later Erian continues limy in the Michigan basin. Inasmuch as in Senecan time terrigenous sediments invade the basin, the history is analogous to that along the Adirondack axis during the Mohawkian epoch, mudstones being restricted by the axis and subsequently passing beyond it. If the Erian influence on sedimentation is indicative of shallower seas over the axis, it may be called synorogenic, for it is synchronous with the advent of the Shickshockian disturbance, but there was not uplift, but lesser depression.

The two sides were synchronously depressed more than the axes only for limited times, and only the northern Adirondack axis had distinct uplift. Secular depression on alternate sides has been considerably responsible for the greater thicknesses on the flanks than on the axes of the present arches.

NORTHWEST TRENDS

Though the dominant trends in the region were northeastward, there were northwest-trending sags through the axes. The Montreal sag northeast of the present Adirondacks is suggested in the distribution of pre-Mohawkian rocks and antedated the formation of the Adirondack arch. Similarly, the Chatham sag between the present Findlay and Algonquin

domes of the Cincinnati axis is suggested in thicker Mohawkian sediments, prior to the formation of the Cincinnati arch. The Frontenac axis at the outlet of Lake Ontario is first suggested in Oriskanian distribution. West of the Cincinnati axis, the Kankakee or Wabash axis has similar trend, separating the Michigan and Illinois basins; it had significance in the distribution of Mississippian sediments (Newcombe, 1933, p. 83). The principal structures of the southern Canadian Shield trend northwest and have been attributed to early Tertiary disturbance (Kay, 1942). Thus trends of some of the oldest and youngest structures cross those dominant in the Paleozoic. Keith (1923, p. 326-327) has discussed the cross folds, but to the writer their alignment is not so striking or significant as Keith believed.

TRANSVERSE MONOCLINES

The transverse monoclinal folds northeast of Lake Ontario (Kay, 1942, p. 632-636) have not been dated or explained. They resemble blocks tilted up to the northwest. The pattern seems en echelon, and suggests relative westward movement on the north.

EFFECTS OF DISTURBANCES IN ADJOINING REGIONS

INTRODUCTORY STATEMENT

Elevation-forming movements to the east of the Allegheny belt produced some or all of the following effects: (1) The uplifts (disturbances) were accompanied by erosion, yielding terrigenous clastic sediments that invaded the region. (2) Warping (epeirogenic) movements within and adjacent to the belt may have been associated with the disturbances. (3) The disturbances, and those in other parts of the earth, may have been accompanied by lowering of sea level (eustatic movements), the withdrawal or shallowing of seas influencing the distribution and facies of the rocks. (4) The loading of the area through any cause may have resulted in warping (epeirogenic) movements of isostatic adjustment. There were also epeirogenic deformations independent of the disturbances in eastern North America. The chronology is presented on the correlation table (Pl. 3).

TIMES OF DISTURBANCE

The character of highlands to the east can be judged by their erosion products. Uplift in middle Trenton (post-Shoreham) time filled the Champlain trough, which had previously received carbonates, with detrital materials; this was the Vermontian disturbance. The abrupt introduction of coarse clastics in eastern Pennsylvania and southeastern New York in late Maysville (post-Fairview) Ordovician indicates a second phase of uplift, the Oswegan disturbance; terrigenous rocks continued

to spread to the west until the close of the period. The Taconian revolution (post-Richmond, pre-Alexandrian) was marked by folding and great overthrusting. Whereas Ordovician sediment progressively coarsened, the Silurian is successively finer and less extensive; they form the "flysche" and "molasse" (Van der Gracht, 1931b, p. 998, 1011) of the Taconian revolution. From estimates of the quantity of material, the volume of rock eroded in the Ordovician was much greater than in the Silurian. The mass raised before the revolution was considerably greater than during the orogeny, particularly when it is considered that some of the Silurian clastics came from rocks raised above base level before the close of the Ordovician. That the elevation was greater preceding orogeny is not a necessary corollary, for there was recurrent uplift before and probably little thereafter.

The next great influx of terrigenous detritus was in the middle Devonian (post-Ulsterian); the later rocks generally coarsen upward and westward, the coarser and nonmarine facies gradationally overlapping the finer and marine. The sediments form the flysche from rising highlands to the east, whose uplift culminated in the Shickshockian (Acadian) disturbance; they are succeeded by Mississippian molasse.

South of the northern Allegheny synclinorium, there is evidence of an earlier Ordovician disturbance, for the Blount sediments of the Appalachian allochthone are thick and terrigenous; they are probably later Chazyan (Ulrich, 1929, p. 49) and came from highlands formed to the east in the Blountian disturbance.

In chronological order, the uplifts that influenced sedimentation in the Allegheny belt were the Blountian (late Chazyan), Vermontian (middle Trenton), Oswegan (post-Fairview Maysville) within the Ordovician, the Taconian at its close, and the Shickshockian (post-Ulsterian) culminating at the close of the Devonian.

CORRELATION OF MOVEMENTS IN THE ALLEGHENY BELT

Movements synchronous with disturbances have been postulated in regions proximal to those of principal displacement; "For these mainly epeirogenic, or moderately orogenic movements during orogenic periods, the term synorogenic has been used" (Stille, 1936a, p. 853). They may be reflected in new patterns of distribution and thickness, the formation or accentuation of troughs and arches.

There is notable change in distribution in the Allegheny belt from the Chazyan to the Mohawkian (Fig. 3d). In the former time, seas from the Champlain trough coursed into the northern part of the belt in the Ottawa Bay; but the Black River introduces the Rideauan trough, the

axis of principal depression shifting from the Champlain trough to the east of the Adirondack axis; it is an example of oscillation. This change may have followed the Blountian disturbance, though the latter is not adequately dated; the Chazyan sediments in the northern Allegheny belt are Upper Chazyan (Valcour) and have been correlated with the Whitesburg and Athens formations of the Blount (Ulrich, 1929, p. 49). If this is correct, the epeirogenic movements in the north followed the introduction of Blountian terrigenous sediments in the southern Champlain trough; and the writer believes that there was similar oscillation along the Adirondack axis in the Appalachians.

Depression on both sides of the Adirondack axis in early Trenton time formed the Adirondack arch. This relative arching preceded the incursion of Vermontian clastic materials. That there was middle Trenton uplift, synchronous with stages of the disturbance, is shown in the truncation of older horizons by the Lacolle conglomerate northeast of the present Adirondack Mountains (Clark and McGerrigle, 1936) and relative uplift by the overlap of basal Denmark in the Mohawk Valley (Fig. 4). The axis of a later Trenton (Cobourg) arch was about 30 miles to the northwest. Movements along the Adirondack belt were not simple nor restricted to the time of Vermontian uplift.

Change in distribution of Cincinnatian sediments in the Adirondack region was synchronous with the Oswegan disturbance. The Lorraine diverges eastward regularly toward the axis and presumably across it; but the Richmond (Queenston) diminishes and disappears toward the southwest Adirondacks. The subsequent block faulting has been interpreted as resulting from "tensional stresses of the period following the Taconic [Taconian] revolution" (Megathlin, 1939, p. 119), though the dating is not conclusive.

The Cincinnati arch is first clearly evidenced in the early Silurian. Albion lithology changes from New York to Michigan, sandy facies being limited to the southeast of the Cincinnati axis; the thicknesses are similar on the axis and its flanks, and the relations are not demonstrative of a relative uplift of a linear belt or arch. In Clinton time, however, calcareous sediment in the Michigan basin was separated from argillaceous beds in the Allegheny belt along part of the axis; there was differential depression of the flanks. Though the first appearance of the arch is approximately synchronous with the Taconian revolution, it is not apparent until after the orogeny.

EFFECTS OF EUSTATIC MOVEMENTS

The demonstration of possible eustatic movements depends on the evidence throughout the earth.

"If it can be shown that transgressions of the sea took place with simultaneous regularity in all the geosynclines of the world, except where modified by local causes, such as minor deformations, excess of sedimentation or other interferences, then the explanation for such processes can only be found in a rise of sea-level." (Grabau, 1938, p. 9.)

The possibility that disturbances are accompanied by widespread marine withdrawals has been recognized.

"If uplift of mountains in some geosynclinal belt is accompanied by crustal changes that for a time somewhat increase the capacity of ocean basins, it is easy to understand that lowered sea-level should result in simultaneous withdrawal of shallow seas from different continental platforms" (Moore, 1935, p. 1256).

There should be marine withdrawal at each known time of disturbance; excessive withdrawal and shallowing on relatively higher areas can be effects of uplift, and seeming lack of retreat and deepening effects of local sinking of greater amount than eustatic fall. Moreover, some withdrawals may be eustatic, but related to disturbances in other regions. Transgressions and regressions are shown on correlation tables (Pl. 3).

The times of Ordovician-Silurian disturbances can be rather closely analyzed and dated, but the phases of the Acadian disturbance of the Devonian are insufficiently known. Because of the proximity of the Allegheny belt to orogenic belts to the east, epeirogenic movements might be more prevalent than in the continental interior, but the proximity permits better correlation of disturbances with regional marine withdrawals.

The oldest Ordovician rocks lie in the northwest Allegheny belt and are lowest Canadian. The subsequent offlap, partial overlap, and offlap of later Canadian units is comparable to that in the interior (Stauffer, 1937, p. 61-62) and suggests eustatic control (Grabau, 1937, p. 735-741). There is no evidence of significant uplift in adjoining regions. The Chazyan sea retreated from this Ottawa Bay, and subsequently the Black River sea advanced. Though the withdrawal may have been synchronous with the Blountian disturbance, the latter seems to have preceded it a little, and there was significant Chazyan epeirogeny in the area. Eustatic fall possibly accompanied the disturbance, but evidence is inadequate.

The Trenton group lies disconformably on the Black River, the break being smallest in the Rideauan trough. Though this relationship is widespread and possibly eustatic in cause, no disturbance of this age is recognized in eastern North America. There was widespread retreat of the sea in middle Trenton (Denmark), somewhat after the Vermontian disturbance began. The younger Cobourg limestone overlaps the middle Trenton along the north margin of the Allegheny belt, and the equivalent Stewartville overlaps in northern Michigan; synchronous withdrawal is suggested

in the Mississippi Valley (Kay, 1937, p. 294). The facies of the Denmark across the Adirondack axis show eastward deepening; hence there was independent epeirogeny. There is disconformity between the Cobourg and Upper Utica southwest of the Adirondacks, probably an effect of local uplift. Thus, the widespread withdrawal in later Denmark time is attributable to eustatic movement accompanying the Vermontian disturbance; movements along the Adirondack axis indicate local epeirogeny.

In northern Michigan and westward, Richmond sediments lie on the Trenton; to the east on Manitoulin Island, Eden and Maysville intervene. It is unknown whether the late Maysville overlaps or offlaps, and thus whether retreat was synchronous with the Oswegan disturbance. If there was eustatic fall in late Maysville, excessive rise followed, for Richmond beds overlie upper Trenton in the Mississippi Valley and northward to the Arctic; the seeming anomalous retreat westward in the Allegheny belt may be attributed to an excess of deposition over eustatic rise and epeirogenic depression.

There is persistent disconformity at the base of the Albion throughout the region, the oldest Silurian being unrepresented in the northern part of the synclinorium. If the Albion be the Castanea sandstone at the top of the Tuscarora in Pennsylvania (Swartz, 1934), the lower and main part of the Tuscarora is older than Silurian on the Cincinnati axis or in the northern margin of the synclinorium. Albion (Brassfield) limestone overlaps on the Cincinnati arch in Kentucky, Ohio, and Indiana (Foerste, 1935), evidencing a somewhat earlier presence of an arch. The pre-Albion (Cataract) disconformity could be due to eustatic movements accompanying the Taconian revolution, but the post-Albion arching is then not synchronous with the same revolution.

There was broad withdrawal in early Lockport, and probably at its close, but no disturbance is known in proximal regions. The later withdrawal of seas between Late Cayugan and Oriskanian might be related to the Caledonian disturbance of this approximate time in Europe.

In summary, there was withdrawal approximately synchronous with each disturbance. It is not known that each is universal, and hence that they are more than suggestively eustatic. The known evidence does not contradict the theory that eustatic fall accompanied each orogeny.

ISOSTATIC ADJUSTMENT

Consideration is next given to the possible effect of isostatic adjustment following loading. The position of deposition depends on other factors. The terrigenous sediments will be carried to areas that are low prior to sedimentation, from those that are high because they had been uplifted—original depression controls the place of deposition (Barrell,

1917, p. 748). Once an initially low area received sediments, the crust would be deformed downward if isostatic adjustment be immediate and local. However, the surface of deposition would rise, for, the mass for unit volume of sediments being less than that of the crust on the order of 5 to 6, there would be depression of the order of 85 feet for each 100 feet laid, and the top would stand some 15 feet above the original elevation. Any relationship in which sediments can be shown to have been laid at successively lower levels requires independent sinking, or eustatic rise. Further applied to the effect of rise of sea level, a change of 100 feet might result in the deposition of several hundred feet of marine sediments, with synchronous depression of the original surface of deposition (Hoffman, 1939, p. 1325). Thus the presence of 1000 feet of marine sediments would not require eustatic rise of more than some 200 feet.

The interpretation that an area has been differentially depressed by loading implies that more material has been laid in one place than another. The distribution of thicknesses and textures will be related generally to distance from and height of producing lands, and in the Allegheny belt clastics diverge toward highlands to the southeast, lines of equal thickness (isopachs) being arcs centered in eastern Pennsylvania. Gradational overlap of coarser and ultimately continental sediments over the marine in the later Ordovician and Devonian is in accord with the requisite rise of surface of deposition of the hypothesis. If loading alone were responsible, the thickness of several thousand feet in each case would require that the surface stood at an elevation of many hundred feet at the close of each period, disregarding eustatic changes, for the basal sediments in each terrigenous pile are of shallow-water origin. But the same result could be produced under any conditions wherein subsidence was slower than deposition, regardless of cause.

Sinking took place both in areas that received and lacked consequential terrigenous sediment. The Canadian and Chazyan carbonates of the Champlain trough approach a mile and lie between quite continuous units of Upper Cambrian and Mohawkian age that become contiguous to the west. The limy sediments of the Michigan basin converge greatly toward the margins and were not in proximity to highlands. The thickening of the higher Silurian (Tonoloway) limestone toward central Pennsylvania is similarly suggestive, though a part of the depression could be attributed to downwarping in areas contiguous to the Bloomsburg red shale delta, which in turn might have loaded the crust. Not only have these areas gone down, but those that received detritus have subsequently sunk. If late Ordovician rocks overlie a plane of deposition that sank because of loading, there was independent subsequent epeirogeny, for

they are overlain by late Silurian and Devonian limestones. That this overlap was not due to eustatic change is shown in the relatively limited extent of equivalent deposits in other parts of the continent.

In nonorogenic times, the adjustments that had preceded are impressive. Despite great contrasts in amount of depression in the Allegheny belt during the late Ordovician, units like the Albion and Lockport are of remarkably uniform thickness and character over great areas. This balance might seem indicative of isostatic adjustment, but, if load controlled depression, surface of deposition should have risen toward areas of greatest thickness. On the other hand, if local sinking directed original deposition, the surface might be built to wave base of the textures represented, finer grades being carried to areas of lower surface of deposition. This should have led to a deficiency of mass in the areas of thickest sedimentation, except that the local sinking might have been induced by previously gained excess.

The conclusion from the record in the Allegheny synclinorium is that local sinking independent of terrigenous loading was a prominent factor in distribution of thicknesses, and that either local sinking complemented uplift, being greatest near the mouths of contributing streams, or that sinking resulted from isostatic adjustment to loading, with subsequent downwarping or erosion to a level in nonorogenic times. The close proximity of this area to the contributing highlands is not conducive to isolation of the factors. The best regions for analysis should be those in which deposits have been carried far from belts of uplift.

SUMMARY

The times of uplift east of the Allegheny belt can be determined from the times of incursion of terrigenous sediments. There were uplifts along axes and sinking of troughs at times of disturbance, but they were not limited to those times. There were marine withdrawals at the time of each disturbance, and these could have been eustatic. Evidence is not convincing that sinking of the crust beneath loaded areas was an effect of isostatic adjustment; and there was significant sinking that cannot be attributed to loading.

LATER PALEOZOIC HISTORY

Post-Devonian Paleozoic sediments lie south and west of the area of study, in Pennsylvania and Michigan; probably they were never deposited along the northern margin of the synclinorium. The lower Mississippian Kinderhookian and Osagian series are represented by the coarse clastic Pocono beds of Pennsylvania. These become finer westward, extending into the Michigan basin, which was separated by the

Kankakee arch from the Illinois basin to the southwest (Newcombe, 1932, p. 78-82). The fine clastic sediments grade into limestone along a belt passing through Illinois (Payne, 1940) and western Kentucky (Butts, 1917; Stockdale, 1939) to eastern Tennessee. The facies patterns are crescentic, with coarsest beds in eastern Pennsylvania nearest the highlands formed in New England in the Devonian. With reduction of the highlands, the earlier Mississippian molasse was succeeded by calcareous sediments. During most of the Meramecan epoch, the Cincinnati arch separated limestones and saline rocks in the Illinois and Michigan basins from limestones in a geosyncline with western margin passing through eastern West Virginia and extending southeastward beyond the limit of preserved rocks of this age. The later Meramecan limestones overlapped into southwestern Pennsylvania (Lucke, 1939) and westward over the Cincinnati arch (Stockdale, 1939).

The pattern of distribution of the later Mississippian and Pennsylvanian marks a significant change in trend from that of the earlier Paleozoic. This is evidenced in the Pennsylvanian Pottsville, which thins rather abruptly from more than 1000 feet to less than 400 feet in a belt passing from southern Pennsylvania through eastern West Virginia, eastern Kentucky, toward northwestern Alabama (Fig. 10). To the southeast, thicknesses exceed or approach a mile (Wanless, 1939; Butts, 1925; Prouty, 1923), whereas there is rather constant thickness of 200 to 400 feet through most of Pennsylvania, Ohio, northern Kentucky, Indiana, and Illinois. This belt of convergence, offlap, and overlap evidences a rapid sinking to the southeast. The structural contrast with its foreland can be compared to that of the middle Trenton between the Champlain trough and the Allegheny belt, though the latter has a much larger proportion of carbonate in the foreland facies. This late Paleozoic belt is designated the Virginian geosyncline; the trough seems to have been southwest-plunging. The western margin crosses that of the Champlain trough of the earlier Paleozoic in southern Pennsylvania and has suggestive relationship to the trend and character of structures formed in the Appalachian revolution. However, the Early Cambrian geosyncline also extended west of the Adirondack axis from Pennsylvania to Alabama.

Later Pennsylvanian and Permian rocks are preserved in diminishing areas in the West Virginia basin on the foreland of the Virginia geosyncline. There is insufficient preservation to permit reconstruction of the paleogeography. They become decreasingly marine and are wholly nonmarine above the Conemaugh (Reger, 1930), evidencing westward withdrawal of shores with the building of the terrigenous deposits.

CLASSIFICATION OF THE ALLEGHENY BELT

APPALACHIAN "GEOSYNCLINES"

The term geosyncline has been applied in the broadest sense to those regions of the earth that have undergone relatively great depression or sinking. Many authors include the Allegheny belt in the Appalachian geosyncline. The stratigraphy of the Paleozoic systems evidences the changing characters of the belts of downward movement and permits their classification.

The deformation of the Allegheny belt is separable into two principal classes of movements—those when terrigenous sediments failed to reach the belt and those when there were highlands to the east whose clastic detritus crossed the Adirondack axis. There was never significant terrigenous materials gained from the west and north; minor quantities are present in the base of overlapping sequences. When terrigenous rocks were absent or quite subordinate within the belt, the margining axes at times formed relatively low arches, as in early and earliest middle Trenton along the Adirondack axis and in Clinton time along the Cincinnati axis. When the troughs bordering the axes were sinking rapidly, the axes became flexures, and the Allegheny belt was relatively depressed along the margins, as in middle Trenton time along the Adirondack, and in Cayugan and Ulsterian times along the Cincinnati.

When there were highlands to the east, the belt was depressed in a semifunneled form, with the apex along its eastern margin, the greatest depression in the area near the terrigenous source and the deformation crossing the Adirondack axis with no appreciable influence. The deformation of the later Ordovician (Cincinnatian) (Fig. 6a) and later Devonian (Erian to Conewangan) (Fig. 10) is of this sort and affects the entire belt, whereas that of the Silurian involved only a part of the breadth. This deformation, attended by thick deposition, in the main has led to the reference of the belt to the Appalachian geosyncline. The belt contrasts with the geosynclines to the east and west, which have undergone great depression irrespective of the character of their sediments; as a whole, the Allegheny belt was never a trough.

The Michigan basin is an elliptical geosyncline (Newcombe, 1933, p. 101). During the Silurian to middle Devonian, it subsided considerably, though terrigenous sediments are practically absent (Pl. 2b). In the later Paleozoic, terrigenous sediments entered the area from the Allegheny belt across the Chatham sag in the Cincinnati arch; but independent subsidence continued during middle Mississippian (Meramecan) when detrital sediments were few.

The Michigan basin seems to have received the centrally thickening sediments because it was sinking, whereas the Allegheny belt was tilted

to a considerable degree because of sinking synchronous with the production of terrigenous sediments to the east, beyond its margin. The former is a geosyncline defined by causes within its area; the latter was depressed in complement to uplift near by. The limits of the former are similar for a long time; the gross sinking of the latter is the sum of periodic movements. Both are within the craton or "shield."

The extracratonic Champlain trough was periodically sinking during earlier Ordovician, when terrigenous sediments were relatively unimportant, its western limit being sinuous. During later Ordovician, when thick terrigenous sediments were laid within, it continued to sink more rapidly than loading alone would cause if isostatic adjustment were immediate and complete. The eastern margin lies beneath the Taconic allochthone but can be interpreted from evidence in the easternmost autochthonous and westernmost allochthonous sections. It has been considered a submarine slope, a flexure with greater subsidence southeastward than in the Champlain belt (Bailey, Collet, and Field, 1928), or as a barrier separating two troughs (Ulrich and Schuchert, 1902, p. 639).

The boulder beds in the northwest marginal Magog trough sediments have been interpreted as submarine landslip deposits by those who believe the Quebec axis was a flexure (Bailey, Collet, and Field, 1928, p. 603-604). The writer concurs that in early Ordovician time there were no highlands along the axis, for, if there had been, there should be thick terrigenous sediments in the Champlain belt; but the presence (Raymond, 1913, p. 30) in local abundance (C. Faessler, personal communication, 1941) of crystalline boulders with exotic blocks of Lower Cambrian to Lower Ordovician limestone (Clark, 1924) such as would be found beneath the Taconic allochthone (Bailey, Collet, and Field, 1928, p. 601-602) require uplift along the axis with southeast sedimentary transfer of the rocks. They may have gained their sedimentary position by landslips but must have been raised before being transported and have been deposited originally in the margin of the Champlain trough. The Cow Head breccia of Newfoundland (Schuchert and Dunbar, 1934, p. 84-86) had similar paleogeographic position and has been attributed to "talus and landslides formed along a fault scarp that came into existence during mid-Ordovician orogeny . . . formed at the nose of thrust sheets"; it has been thought to be autochthonous. The post-lower Trenton Rysedorph conglomerate in the Magog belt southeast of Albany (Ruedemann, 1901; 1930) contains exotic boulders of Champlain trough facies (Kay, 1937, p. 276-277) and has been attributed to the "erosion of anticlinal ridges" (Ruedemann, 1930, p. 113).

The Quebec axis probably did not separate faunas, but facies, which in turn influenced faunas. There were uplifts along the axis, some of which may have risen above sea level, any of which must have directed currents and thus controlled distribution of facies. The conglomerates at Quebec were formed by movements antedating the Vermontian disturbance, for they are in Canadian beds and evidence the prolonged activity along the axis. The comparable "exotic" blocks in the northern Ouachita allochthone have been attributed to thrusting of rocks toward the geosynclinal margin (Hendricks, 1940); or landslipping from an archipelago, possibly bordered by normal faults (Kramer, 1933; Moore, 1934). The Haymond conglomerate of west Texas has boulders supposedly "derived from the erosion of thrust sheets advancing from the southeast" (King, 1937, p. 91), that is from the Ouachita geosyncline; their deposition antedates the important Marathon orogeny and thrusting. Those in the Ordovician Woods Hollow formation of the same region (King, 1937, p. 24-25) are of foreland facies. These are analogous to those along the Quebec axis.

A structurally active and relatively higher though perhaps interrupted barrier formed the eastern margin of the Champlain trough. To the west lay the Adirondack belt, alternately an arch and flexure. The trough sank endogenetically during Cambrian and earlier Ordovician time and became deeply depressed as it filled with terrigenous sediments from the Vermontian geanticline in the western Magog belt in middle Trenton.

CLASSIFICATION OF GEOSYNCLINES

Schuchert (1923, p. 195-197) has classed troughs such as the Champlain and Magog as monogeosynclines, formed by the partition of the earlier polygeosyncline, the "St. Lawrence Trough," by an intrageosynclinal geanticline, the Quebec "barrier." Stille (1936a, p. 517) has separated geosynclines into the greater intercontinental belts, orthogeosynclines, and those lying within the shield areas or cratons, "parageosynclines," the latter a term having a different earlier definition (Schuchert, 1923, p. 199). Inasmuch as the character of areas is transitory, any classification is fundamentally temporal. The writer adopts an Ordovician basis of reference for definition of paleogeographic belts; this may reflect his interest in the system, but it seems advantageous because the record is extensive, the belts had become distinct, and had not been altered by orogenies.

The Magog geosyncline has graptolitic slates, formed in at least moderate depth, and associated cherts and volcanic rocks, as in northern Newfoundland (Sampson, 1923; Heyl, 1936) and southeastern Quebec (Cooke, 1935; Clark and Fairbairn, 1936); they are in a belt later

intruded (Keith, 1928). Lower Ordovician graptolites are of distinctive genera, and practically limited to the shaly and cherty sequences of the Magog belt, reflecting a structurally controlled facies. Similar rocks in the Ouachita area contrast with carbonates on the foreland, and they are found in the western belt of the "Cordilleran geosyncline" (Ferguson, 1924, p. 20; Ross, 1934, p. 942-945). That facies rather than land barriers limited their distribution is shown by thin graptolitic shales intercalated within the carbonates in Oklahoma (Decker, 1936), West Texas (King, 1937, p. 26-30), Utah (Clark, 1935), and southeastern Idaho (Mansfield, 1927); that local conditions changed, by the intercalation of thick graptolitic shales within limestone formations, as in British Columbia (Walker, 1926, p. 24-31; Evans, 1933).

The Magog belt lies in a pliomagmatic zone (Stille, 1936a), and the trough is a eugeosyncline.[3] Eugeosynclines are characterized by a prevalence of shales and cherts, the presence of volcanic rocks, ophiolites or greenstones (Stille, 1936a, p. 517; Bailey, 1936, p. 1717-1723); the rocks have been interpreted as formed at great depths (Ruedemann, 1936, p. 1545-1563) to an extreme that does not seem warranted by the coarser texture of some of the interbeds. They have been the locus of principal subsequent plutonic invasions.

In contrast to the Magog eugeosyncline, the Champlain belt contains dominant carbonates of shallow-water origin, unaffected by subsequent volcanism; it is a miogeosyncline. The lithologies are more like those on the craton than in the eugeosyncline, and with the craton, the belt is essentially amagmatic. But the miogeosyncline is extracratonic, inasmuch as the sediments in most regions have been thrust subsequently on the foreland.

Just as the Magog and Champlain troughs comprise the two parts of the St. Lawrence orthogeosyncline,[4] there are two similar stratigraphic belts in the Cordilleran orthogeosyncline. The western, eugeosynclinal belt has been classed as pliomagmatic (Stille, 1936b, p. 139-143) and called "Nevadan." Inasmuch as the name Nevadian has been applied to the intrusive disturbance at the close of the Jurassic period, it is

[3] In reply to the writer's letters of November 8, 1939 and April 23, 1941, in which he requested a definition of orthogeosyncline and emphasized the contrasts between the stratigraphy of the pliomagmatic and amagmatic zones of Stille and the similarities of the stratigraphy of the latter with that of the craton, Stille, after discussion in a letter of March 20, 1940, proposed the following classification in a letter of May 24, 1941:

"1. Orthogeosynklinalen (alpinotype Geosynklinale) zerfallend in
a) eugeosynklinale Zonen (pliomagmatische Mutterzonen der Interniden)
b) miogeosynklinale Zonen (miomagmatische Mutterzonen der Externiden).
"2. Parageosynklinalen (germanotype Geosynklinalen auf schon konsolidiertem Untergrunde)."

The terms eugeosyncline and miogeosyncline have been defined recently (Stille, 1941, p. 15).

[4] An orthogeosyncline is defined on the character of the rocks and their history and is preferred to polygeosyncline, which is a geometrically descriptive term. Orthogeosynclines are polygeosynclines, but not all polygeosynclines are orthogeosynclines.

confusing to use Nevadian or Nevadan for the stratigraphic belt. The lower Ordovician (Deepkill) graptolitic shales in Idaho (Umpleby, Westgate, and Ross, 1930; Ross, 1934) may be in the belt; equivalent rocks in southeastern Alaska (Buddington and Chapin, 1929, p. 72-79) are typically eugeosynclinal, and volcanic and cherty slates are prevalent in the late Paleozoic, Triassic, and Jurassic of western British Columbia. The eastern, miogeosynclinal belt has the well-described sequence of southeastern Idaho (Mansfield, 1927) and is the Cordilleran geosyncline of most authors. Keith (1928) emphasized the importance of subsequent plutonic intrusions in the western belt; Stille illustrates (1936b) the movements along the eastern border as westward, but the belt was thrust on the eastern belt (Umpleby, Westgate, and Ross, 1930, map; Merriam, 1940, Pl. 1), much as the Magog was thrust on the Champlain. The overthrust sediments of the eugeosynclinal belt form an unnamed allochthone inasmuch as the orogeny that produced the structure has not been dated; it is certainly post-Carboniferous and probably pre-Nevadian. Nolan (1929) has described the geosynclinal nature of the eastern belt in Nevada.

The Allegheny belt seems comparable to the Coloradan, to the east of the Cordilleran belts, each having had at some times depressions filled with detrital materials gained from outside the belt; it is proposed that these intracratonic, terrigene-filled depressions be called deltageosynclines (Fig. 6; Pl. 2), in contrast to the endogenetic geosynclines like that in the Michigan basin that are termed autogeosynclines (Pl. 2b) in reference to their independent sinking and isolation from those near the more active tectonic zones.

In eastern North America, the sequence has been development of the orthogeosynclines, uplift of an intraeugeosynclinal geanticline Vermontia, folding and thrust of the eugeosynclinal Magog belt on the miogeosynclinal Champlain (Taconian revolution), intrusion within the eugeosynclinal (Acadian disturbance) and folding and thrust of the miogeosynclinal Champlain on its foreland, the Allegheny belt of the craton (Appalachian revolution). A comparable sequence seems represented in the Cordilleran belts in the Mesozoic and early Tertiary history. In the Ouachita region, magmatic invasions are not recognized, though the observable part of the belt is quite marginal; the thrusting is not dated with certainty, though it is at least in part intra-Pennsylvanian, and two belts of thrusting are not known. Exotic boulder conglomerates are present on the foreland margin of eugeosynclines—the source in the miogeosynclinal belts—and the deposits preceded the major thrusting. In a discussion of similar stratigraphy in the Ural region of eastern

Russia the writer (Kay, 1941b) misapplied the term orthogeosyncline to the more restricted eugeosynclinal belt.

The Ouachita and Virginian belts illustrate the confusion if one does not recognize the temporal nature of the belts. If one grant that the earlier Paleozoic rocks of the former are eugeosynclinal, though they lack characteristic volcanics, they are overlain by sediments of the Pushmataha series (Harlton, 1938, p. 853) that are like and presumably continuous with the lower Pennsylvanian sediments of the Virginian geosyncline (Van der Gracht, 1931) that is marginally intracratonic with reference to the earlier Paleozoic definition but is miogeosynclinal. On the other hand, the early Paleozoic "Ouachita" eugeosyncline must trend southeastward from Arkansas if it joins the extension of the Magog belt. It may be that the Pushmataha geosyncline is related to a different cause from the earlier trough and that the "Ouachita" geosyncline is composite. The great thrusting in the Virginian-Pushmataha belt may be limited to the area in which it lies on eugeosynclinal facies. That the Virginian miogeosyncline trended diagonally to the Ordovician Champlain trough has been stated (Fig. 10); whatever factors determined the trend and character of early Paleozoic belts had been superseded by others in late Paleozoic time.

Belts that had been sinking independent of terrigenous loading sank more when orogenic events produced quantities of detritus in neighboring belts. The Champlain trough developed this contrast in middle Trenton but retained its linear limits; and the later, transverse Virginian trough likewise maintained an essentially linear northwest boundary whether terrigenous rocks were unimportant or preponderant. These miogeosynclines contrast with the great semilenticular depressions of the simple deltageosyncline, or the more sinuous and varied outlines when the terrigenous sediments were distributed from a number of centers in other deltageosynclinal belts.

Foreland folds within the eastern border of the Allegheny belt failed to gain the magnitude of those of the comparable Coloradan belt of the late Paleozoic and particularly the Cretaceous; even during the late Gulfian epoch, uplifts within the Coloradan deltageosynclinal belt yielded significant quantities of clastic sediment (Knight, 1938); nor were they as important as the uplifts along the Wichita-Amarillo axis in the Ouachita foreland.

BARRIERS AND TROUGHS, AND OSCILLATION

Though the concept of barriers and troughs has been credited to Logan (1866, p. 45), the comprehensive application of the theory to the North American stratigraphy has been that of Ulrich and Schuchert (1902).

"Though abundant corroborative evidence of the existence of a narrow barrier between the stratigraphically inharmonious areas is afforded by the structural geology, . . . the striking dissimilarity marking the faunas pertaining to the lithologically equally dissimilar Ordovicic rocks lying respectively on the east and west sides of the Great Valley emphasized the separate provinces" (p. 634).

The theory was elaborated by Ulrich (1911), who emphasized the "permanence of certain continental lines of elevation separating areas of depression" (p. 292) and introduced the principle of oscillation, "that the deposits on the opposite sides of the Appalachian barriers are as a rule not contemporaneous but successive" (1911, p. 292; 1929, p. 55). The barriers rather than the facies were considered directly responsible for the contrasting faunas (1911, p. 512-513);

"correlations requiring the assumption of lateral changes in faunas caused by differences in character of bottom . . . are commonly found to be in error" (1916, p. 454). ". . . owing to the overthrusting of the middle and eastern troughs in the Appalachian Valley, . . . the eastern edges of their respective formations are as a rule inaccessibly buried" (1911, p. 452).

The structures of the Appalachian revolution being transverse to the Ordovician, the relatively undisturbed area of the northern Allegheny synclinorium eastward reveals relations that may explain less evident relations in the thrust-separated sequences of the southern Appalachian Mountains. The character and influence of the Adirondack axis as a "barrier" has been studied, and the basis for the interpretation of the Quebec "barrier." The stratigraphic contrasts on the two sides of the Saltville-Helena thrust being comparable to those on the flanks of the Adirondack axis, and the latter being traceable toward the termination of the thrust, there are grounds for a southern "Rome Barrier" (Ulrich and Schuchert, 1902, p. 644) or "Cahaba Barrier" (Butts, 1927, p. 16), but only during Ordovician time; the principal contrasts across the thrust would be attributed to great displacement (Van der Gracht, 1931a, p. 111).

The writer interprets the "barriers" as having influenced facies through controlling distribution of sediment and attributes many of the faunal contrasts to the facies and original geographic separation. Thus, Ulrich considered that the Athens shale was laid in the "Athens Trough," continuous with the "Levis Trough" in which the homotaxial Normanskill was laid; but the former was in the southern continuation of the Champlain trough, and the latter in the Magog to the east, the Blountian disturbance having led to the deposition of sediments in the south comparable to those in the Magog in the north. Five troughs were postulated between the Adirondack and Green mountains of New York-Vermont (Ulrich, 1911, p. 442-446); the rocks of the third are those of the Taconic klippe, those of the first, second, and fourth are continuous in field outcrop (Cady, 1938), and the fifth includes meta-

morphic rocks considered (Keith, 1932, p. 404-406) to be of the same sequence as the third. As now understood, the first, second, and fourth are parts of the continuous section of the Champlain trough, and the third and fifth were laid in the Magog.

The idea of multiplicity of troughs seems to have been developed from two independent considerations. First, having concluded that some thrusts are along barriers, it was assumed that most or all are along barriers, and that each thrust slice corresponded to an original trough. Second, in the belief that the barriers persisted through the Paleozoic, each barrier might be based on evidence from a single time and interpolated into the history of all periods. It is now thought that the later troughs had independent limits and more westerly trend than those of the middle Ordovician. The "transverse axes" seemed evidenced by related phenomena; inasmuch as the later troughs cross the earlier, Lower Devonian sediments, for example, are present in the Champlain belt in the north but absent in the south because the Devonian trough crossed the Adirondack axis, by then dormant, into the Allegheny belt. The disappearance of units is further related to thrusting. Just as the rocks of the Champlain belt pass beneath the Taconic allochthone in Quebec (Bailey, Collet, and Field, 1928, p. 602-604) and New York, so also those of the Champlain belt in the western slices in northern Virginia pass beneath more easterly thrust sheets south of Roanoke, and the eastern Allegheny belt sediments beneath the Appalachian allochthone in the southern Appalachians. Moreover, the Champlain belt sediments may have been eroded off the more easterly thrust sheets southward inasmuch as the belt passes from the early Devonian trough in the north to the region east of the trough in the south.

Ulrich's (1911, p. 415-416) evidence for oscillation for the Adirondack area is principally correct (Fig. 4), and there seems comparable foundation for the Chambersburg example (p. 321-329). These may be more exceptional than he believed, however. Some of the evidence that seemed significant is related to the changing trends of later troughs.

In summary, the writer believes that there were a few barriers and troughs in the Ordovician and that faunas were controlled and limited by facies, which in turn were dependent on fundamental structural changes. The middle Ordovician troughs did not persist long, but others, of more southwesterly trend, followed. The Appalachian structures seem directed by their presence or by the more fundamental causes of their origins. Oscillation is represented in the stratigraphy but is not thought to have been the principal cause of contrasts. Appalachian thrusts have a major part in the distributions now apparent, just as have the earlier Taconian thrusts.

APPALACHIAN REVOLUTION

THE FOLDS

The Appalachian revolution came at or near the close of the Paleozoic era. The more obvious and spectacular effects are in the Appalachian Mountains from eastern Pennsylvania to Alabama, but folds of some magnitude extend into the Allegheny synclinorium of northwestern Pennsylvania and southern New York. The trends are northeast in western Pennsylvania (Sherill, 1934; Cathcart, 1934; Fettke, 1938, p. 152-154), becoming nearly east in southern New York (Wedel, 1931; Fox, 1932). The northwest limit of folds having structure of more than 100 feet passes through southern Cattaraugus, Allegany, and Livingston counties, New York; to the north, the regional structure is essentially homoclinal. The asymmetry of northern folds contrasts with that in the more strongly folded Appalachians in that the steeper flanks face southeastward rather than northwestward; the significance has been discussed (Sherrill, 1934). The trend of Appalachian folds in New York departs by nearly 45° from that of the Adirondack axis, the divergence decreasing southward.

The northern Appalachian foreland folds parallel the Appalachian Structural Front (Price, 1931). The front lies west of the Virginian geosynclinal margin in the south, and far to the northwest in Pennsylvania (Fig. 10). It does not follow any recognized axis, flexure, or trough of the Paleozoic, and the origin of its specific position is not evident. It does lie in the position of thick Paleozoic sediments, but these vary in manner of deposition, the principal thickness in the north being in the Ordovician and Devonian deltageosynclines and in the south in the Virginian miogeosyncline. The generalization is not definitive of the line of contrast at the front.

THRUST FAULTS

The folded and faulted western Appalachians lie in the Allegheny belt; faults of considerable horizontal displacement pass into folds at their ends. The Saltville and Helena faults separate middle Ordovician rocks comparable to those in the Champlain and Allegheny belts farther north; and the northern termination of the Saltville thrust is about on the extension of the Adirondack axis. Though stratigraphic contrasts are small across the Saltville thrust as it approaches its termination in Virginia, they are great across the Helena thrust in Alabama (Butts, 1927, p. 16); the contrast in Alabama has been attributed to the relatively great displacement (Van der Gracht, 1931a, p. 111-112). The Saltville-Helena thrusts form the margin of the Appalachian allochthone, in the

sense that they separate two stratigraphic provinces, of which the margin of the eastern has been thrust on that of the western. The allochthonous rocks of the Saltville slice pass northeastward into the Champlain belt

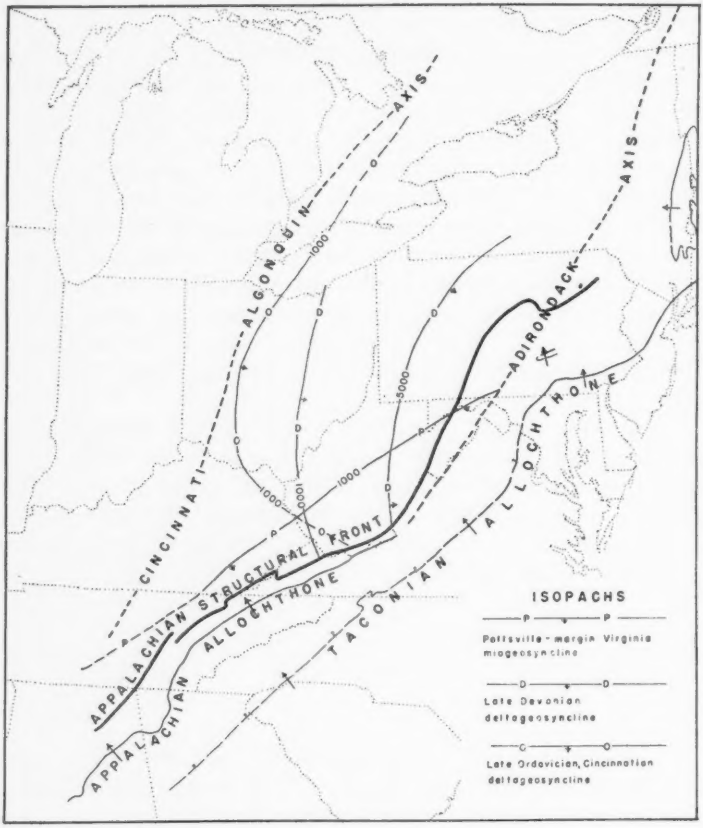


FIGURE 10.—*Appalachian structural front with reference to Paleozoic geosynclines*

where they are autochthonous with respect to the Taconian thrusts, and rocks eastward to the Martic thrust were laid in the Champlain belt. Southeasterly fault lines extend progressively farther northeastward in Virginia (Butts, 1933, map), comparably to the southward termination of the principal thrusts in the Champlain belt of western Vermont (Cady, 1938).

The western fault lines of the Appalachian system are similar to the margin of the Virginian geosyncline in position and trend, but those farther southeast increase in throw southwestward, and thus the rocks were originally farther east in the south; the trend of the original breaks was more northerly than the present fault lines. Relatively great displacement toward the south has rotated the slices so that the fault lines almost parallel the margin of the Virginian belt, but the trend of the initial fractures must have been more southerly. The thrust along the margin of the Appalachian allochthone developed along the Adirondack axis. This axis parallels the Cincinnati axis in the north, the two lying 300 miles apart from New York to Virginia. If the parallelism continued southward, the displacement on and west of the Helena thrust in Alabama would be 100 miles or more. The amount seems unreasonably large, the two axes were structurally active at different times, and their parallelism may be coincidental.

Keith has stated that

> "identity of the belt of overthrusts with the belt underlain by thick sections of weak Cambrian or Algonkian beds and floored by a . . . slope of the pre-Cambrian granite and gneiss is so complete that they must be acknowledged as cause and effect" (1928, p. 346).

Considering the continent as a whole, this generalization is warranted, in spite of the fact that the extent and western margin of the earliest Paleozoic Cambrian "geosyncline" in the Appalachian region is unknown but is far to the west of the zone of the principal thrusts; nor is the extent of the pre-Cambrian geosyncline known. The pre-Paleozoic rocks in the Taconic and Martic thrust sheets are metasediments where exposed, but unfortunately the limits are poorly defined westward. There are known areas of granitoid rocks in the Champlain belt in the north; the Taconian thrust sheets developed in the eugeosyncline, and possibly the pre-Cambrian metasediments were laid in the same belt; the longitudinal extent of the Glenarm series of metasediments and their essential parallelism to the Taconian thrust lines are certainly suggestive.

Ashley has written that "apparently the great thrust moved somewhat crosswise to the original lines of sedimentation, which extended more nearly north and south, as suggested by the [Devonian] isopachous lines" (1938, p. 426). Inasmuch as the latter have arcuate pattern when considered as a whole and reflect semilenticular depression, the divergence in trend varies with points of reference; but the isopachs and the Appalachian Structural Front in Pennsylvania strikingly approach parallelism (Fig. 10). The front roughly parallels the margin of the Virginian belt in the south and the deltageosynclinal isopachs in the north.

The line of the eugeosynclinal margin corresponds to that of initiation of the Taconian thrusts whatever the reason for the specific position;

and the margin of the Champlain belt corresponds to that of the thrusts on the margin of the Appalachian allochthone. The cause would seem more fundamental than the character of the surficial rocks; their nature may reflect the same basic structural conditions. Similarly, the principal

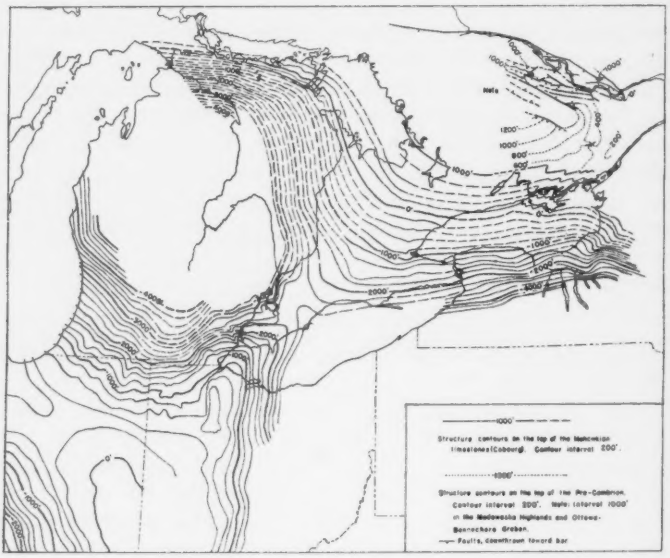
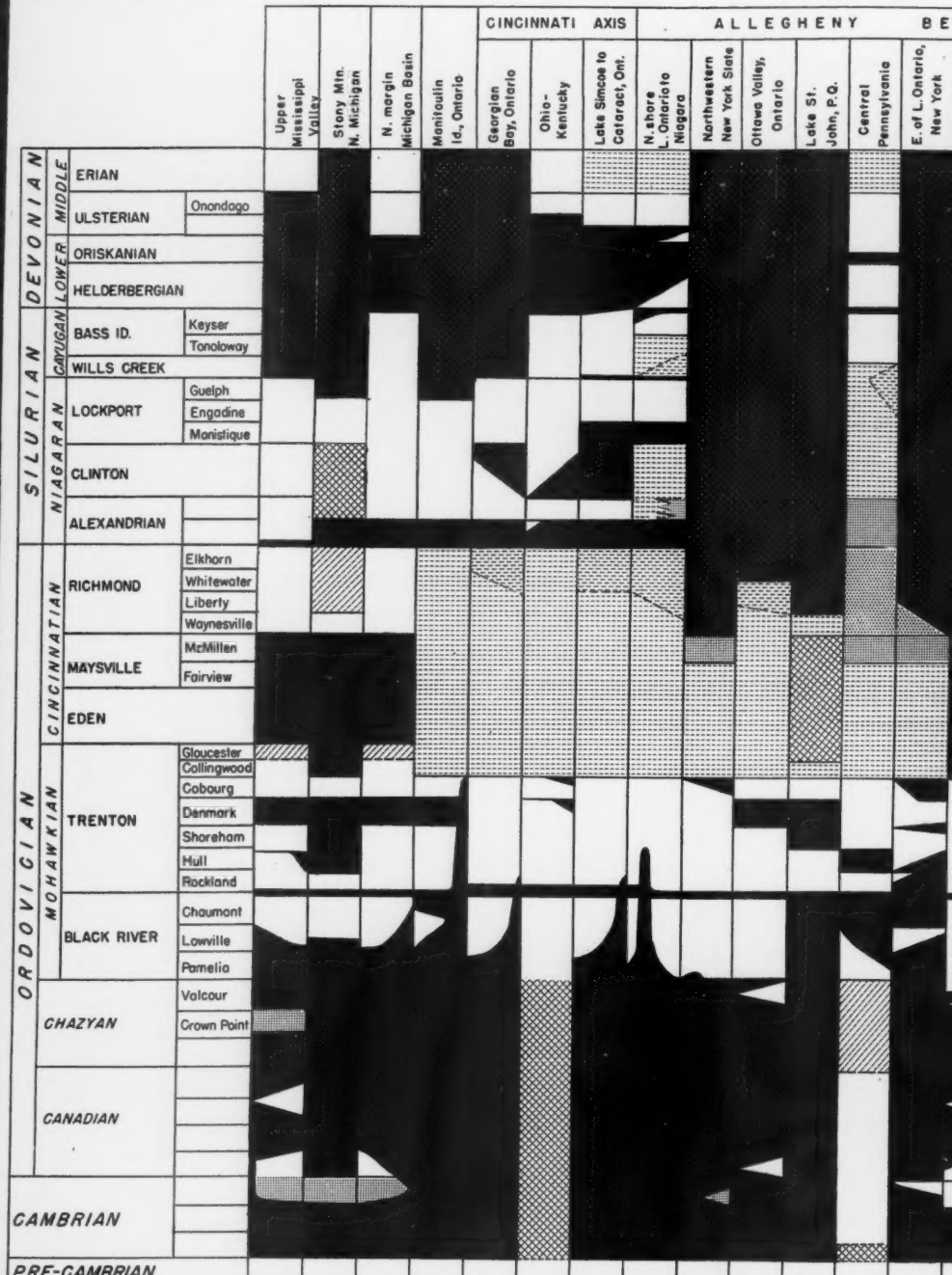


FIGURE 11.—*Structure contours on top of Mohawkian limestone*

Laramide thrusts formed within the eastern margin of the Cordilleran miogeosyncline (Keith, 1928, p. 333-334), and others, perhaps older, along its western margin (Merriam, 1940), east of the eugeosyncline.

LATER DEVELOPMENTS OF THE NORTHERN ALLEGHENY SYNCLINORIUM

The minor transverse monoclinal folds within the northern margin of the synclinorium (Kay, 1942, p. 632) are of unknown age and origin. The border is essentially homoclinal north to the uplifted fault block on the south side of the Madawaska Highlands of Ontario. The region to the north has been extensively block-faulted by movements attributed to the close of the Mesozoic era or the early Tertiary. Subsequent denudation implies Tertiary uplift of the whole region. The present state is shown in a structure contour map on the top of the Mohawkian limestones (Fig. 11).



CORRELATION OF PRE-MIDDLE DEVONIA

E DEVONIAN SEDIMENTS

SUMMARY

The Allegheny synclinorium is a structural belt of complex origin. It was first defined on the east by the formation of the Adirondack axis in the Ordovician and on the west by the initiation of the Cincinnati axis in early Silurian. The great thickness of sediment is principally an effect of its position in the northern quadrant of great semilenses of terrigenous rocks laid in Ordovician, Silurian, and Devonian deltageosynclines; otherwise it stood relatively higher than the Champlain trough and the Michigan basin on the two sides. The folds of the Appalachian revolution invade it in the south, and the southern synclinorium is partially buried beneath the Appalachian allochthone, the bordering fault having originated along the original southeastern margin of the Allegheny belt. Though the Frontenac axis in the northeast seems of earlier Devonian origin, there was significant block faulting along the northern margin in late geologic time.

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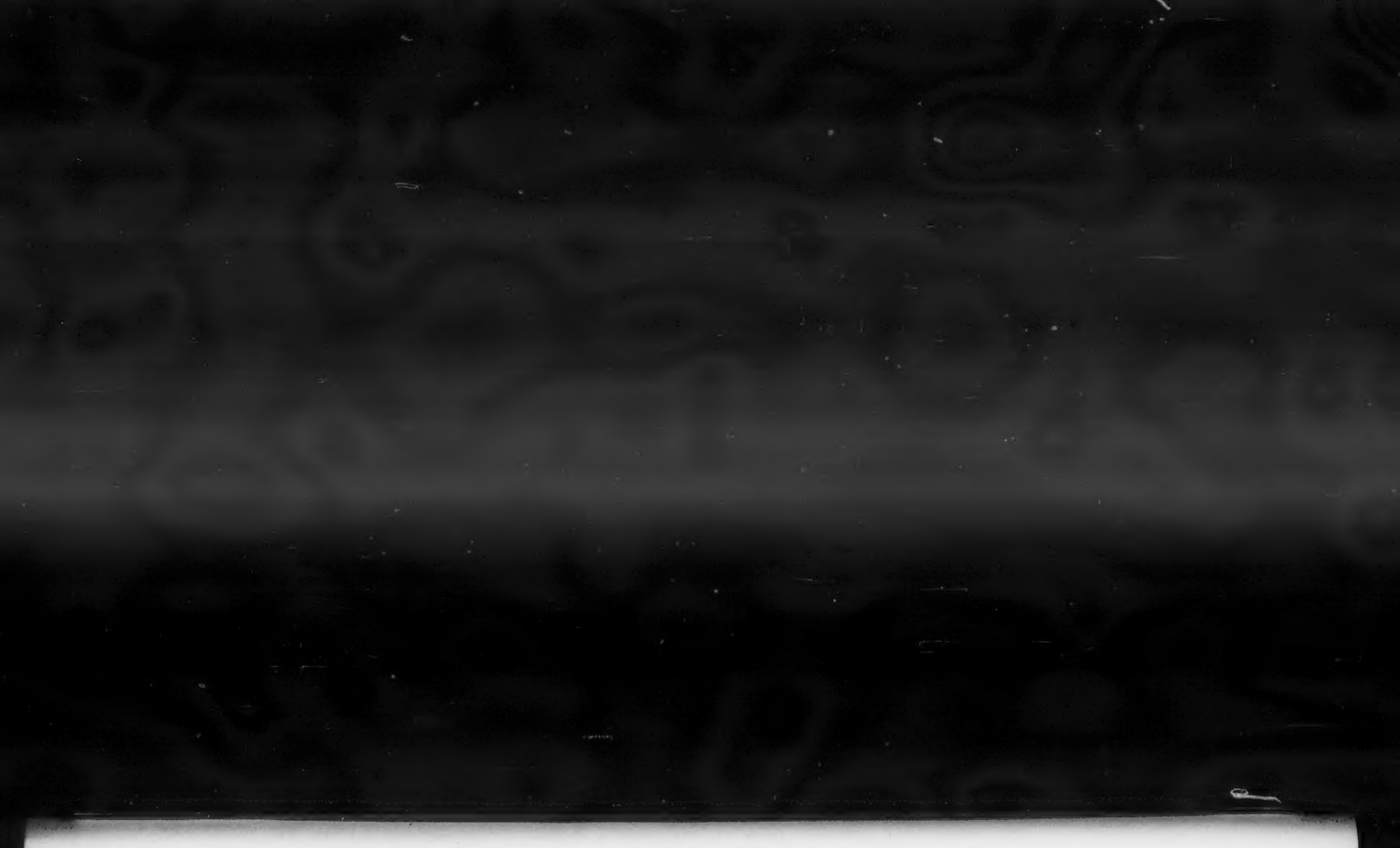

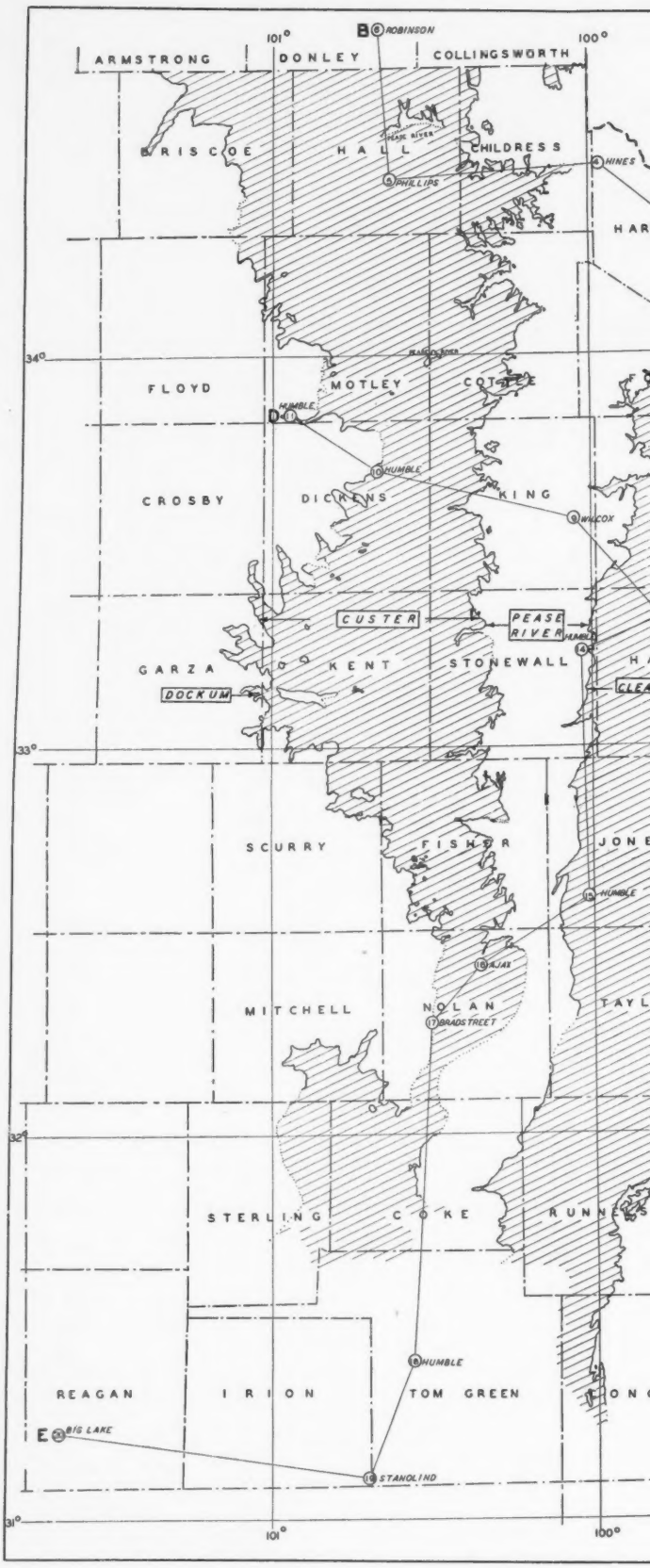

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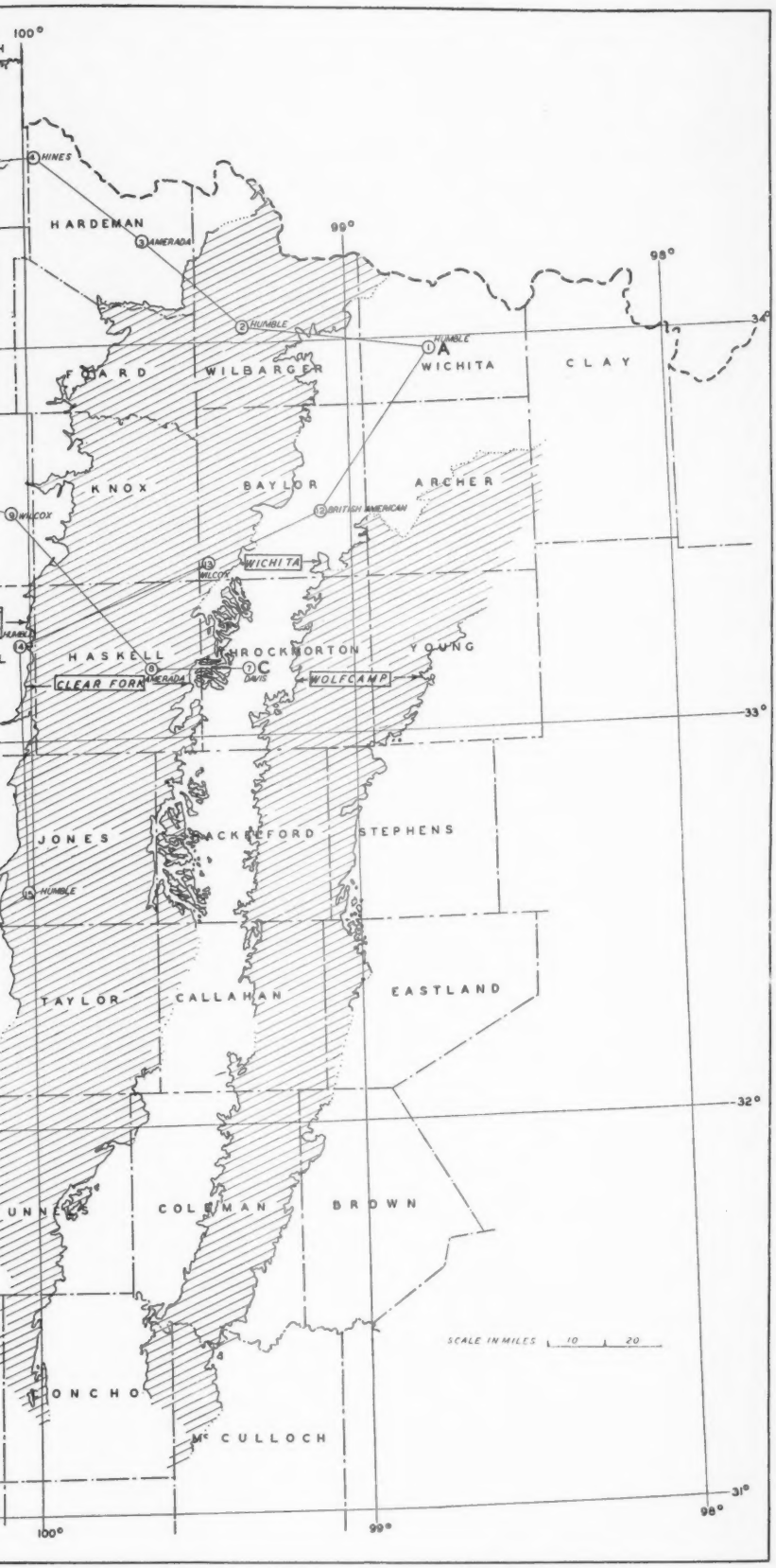

BASIN INCLUDING SMALL PORTION OF THE PERMIAN BASIN

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WEST TEXAS BARRED BASIN

BY ROBERT I. ROTH

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ABSTRACT

The regional stratigraphy of the Leonard series along the eastern portion of the Permian Basin of West Texas is set forth by means of a number of strategically

placed stratigraphic sections. Important faunal zones are indicated and discussed. Many characteristics of a barred basin environment are reviewed, and the hypothesis is advanced that the Leonard was deposited in a barred basin.

ACKNOWLEDGMENTS

The writer began surface and subsurface studies of the Permian Basin in 1929 while working for the Indian Territory Illuminating Oil Company of Bartlesville, Oklahoma. This company gave every possible aid, which is gratefully acknowledged. Mr. John W. Skinner has identified or checked the identification of all fusulinids found. Without this help regional correlations would be difficult. Drs. A. K. Miller, R. C. Moore, and P. B. King have read the manuscript and offered many valuable criticisms. Finally, all help necessary to the further study of the problem and permission to publish were given by the Humble Oil & Refining Company of Houston, Texas.

INTRODUCTION

The peculiar sedimentary record preserved in the Permian Basin of West Texas has elicited many explanations both for the environment and the source of these sediments.

In the Glass Mountains, the Leonard limestones, sandstones, conglomerates, black shales, and bioherm masses represent a normal though shallow marine environment. To the northeast and north they give way to black shales. Continuing along the above directions these shales grade rapidly into beds of limestone, dolomite, anhydrite, sodium chloride, reddish brown to greenish gray shales, and grits. Certain horizons show minor sandstones and conglomerates. The composite sedimentary aspect compares favorably with those of barred basins and indicates a similar environment.

ZONE OF *PERRINITES* OR TYPE LEONARD

The zone of the ammonoid *Perrinites* is defined by Miller (1938, p. 1014-1019) to comprise beds belonging to the Middle Permian Leonard series (Fig. 1). Sediments of the zone represent almost every type of environment and facies—normal marine, euxinic, and desiccated—and no known barrier separates the several types.

The Leonard, recently elevated to the rank of series, is one of the stratigraphic units of West Texas which were defined and named long before their faunal content was clearly understood. At Leonard Mountain, its type locality, it is bounded by an unconformity at the base and a disconformity at the top. Faunally it is characterized by *Perrinites* and *Dictyoclostus bassi* McKee.

Fourteen miles northeast of Leonard Mountain a wedge approximately 550 to 600 feet thick is included at the base of the Leonard because there is no obvious break at its top. This wedge, not present in the type area, is characterized by *Schwagerina crassitectoria* Dunbar and Skinner.

	TEXAS				
PERMIAN	AMMONOID ZONES	GUADALUPE AND DELAWARE MOUNTAINS	GLASS MOUNTAINS	NORTH CENTRAL TEXAS	FUSULINID ZONES
	CYCLOLOBUS	XXX	XXX	XXX	
	TIMORITES	CAPITAN UPPER DELAWARE	CAPITAN	CUSTER IN PART	POLYDIEXODINA
	WAAGENOCERAS	MIDDLE DELAWARE LOWER DELAWARE	WORD	XXX	PARAFUSULINA
	PERRINITES (BY DEFINITION)	BONE SPRING	LEONARD	PEASE RIVER CLEAR FORK	
		X	XX	WICHITA	SCHWAGERINA *CRASSITECTORIA*
	PROPERRINITES	WOLFCAMP X	WOLFCAMP	WOLFCAMP	PSEUDOSCHWAGERINA
PENN.	UDDENITES	MAGDALINA X	GAPTANK	CISCO	TRITICITES
	X NOT EXPOSED. XX PRESENT. XXX ABSENT				

FIGURE 1.—*Permian ammonoid-fusulinid zones (Texas)*

Neither *Perrinites, Properrinites,* nor *Dictyoclostus bassi* McKee have been reported from it. In boring No. 19 in Irion County, 175 miles northeast of Leonard Mountain (Pl. 1), *S. crassitectoria* has been found to range through an interval of about 600 feet. In boring No. 18 in Tom Green County, 23 miles north and a little east of No. 19, the same interval is 620 feet. Apparently the *S. crassitectoria* zone maintains a remarkably uniform thickness over great areas. *S. crassitectoria* has not been reported from any well north and east of boring No. 18.

On the east side of the Permian Basin about 600 feet of strata lying above the Wolfcamp, as now defined, contains the ammonoid *Proper-*

rinites (Fig. 1). These strata, which are called the Wichita group, may be traced lithologically from well-to-well southwestward into the zone of *S. crassitectoria*. Therefore, the lower part of the Leonard series, as defined in some areas, but seemingly not in the type section, may be older than the zone of *Perrinites* and may contain *Properrinites*. If this should prove to be the case, the zone of *Perrinites* is not synonymous with the Leonard series as this term is now being used. Proof of the supposition just offered is lacking, however, and the author follows Miller's definition of the zone of *Perrinites*.

ECONOMIC ASPECTS OF THE LEONARD

The importance of the economic aspects of the Leonard cannot be over-emphasized for a large percentage of its commercial oil and gas seems to have originated therein. Immediately preceding the deposition of the Leonard a broad emergence developed continuous porosity by weathering and alteration of the underlying limestones into dolomites and porous limestones that now serve as reservoirs. Very little oil or gas has been discovered in that portion of the Leonard underlying the San Angelo horizon. On the east side of the Permian Basin proper a few small areas have yielded oil in commercial quantities from strata above the base of the San Angelo. One of these producing areas is in Garza County. Over most of the Permian Basin, pre-Custer weathering developed continuous porosity and dolomitization in the limestones beneath the Custer. These magnesian limestones, some of which are younger than Leonard, have been loosely referred to as the West Texas big lime. Locally porosity has developed through a stratigraphic interval of 300 feet or more and in many places this porosity is filled with oil. Very little oil or gas has been found in the areas where the big lime grades into strata which are predominantly chemical deposits such as sodium chloride, anhydrite, and primary dolomite.

MAJOR SUBDIVISIONS

GENERAL STATEMENT

On the east side of the Permian Basin, the Leonard may be subdivided into the Wichita, Clear Fork, and Pease River groups (Pl. 1). Their characteristics, both on the surface and in the subsurface, are as follows, starting with the oldest.

WICHITA GROUP

The beds from the top of the Elm Creek limestone upward to the top of the Lueders limestone are assigned to the Wichita group. A regional

unconformity marks the base of the group, and probably there is a disconformity at the top. The average thickness is 675 feet. In the type area, the group consists of red clastics which are very difficult to separate from redbeds of the overlying Clear Fork and underlying Cisco. The topography of the Wichita redbeds is rolling and broken by many local badlands; that of the marine Wichita beds is rough on account of strong development of stairlike escarpments.

In the subsurface, the Wichita grades from red clastics into evaporites and marine sediments. At or near the base is a widespread zone of anhydrite. Where the Clear Fork becomes marine, the top of the underlying Wichita is difficult to recognize for in these areas its deposits are also marine.

CLEAR FORK GROUP

This group includes all beds from the top of the Lueders limestone up to the base of the San Angelo sandstone. Disconformities probably occur both at the top and bottom. The average thickness is 1150 feet. Except for a few magnesian limestones, which are exposed in the southern area, practically the entire section consists of soft reddish-brown clastics. A few beds of gypsum and anhydrite are exposed in the upper portion. The Clear Fork outcrop is a monotonous plain interrupted locally by badlands.

In the subsurface the Clear Fork red clastics grade westward into marine sediments and evaporites. Where the San Angelo changes to dolomites and marine sediments, the top of the Clear Fork is difficult to recognize because in these areas it also is marine.

PEASE RIVER GROUP

The rocks of Leonard age that overlie the Clear Fork group unfortunately have been designated as the San Andres group, the El Reno group, the Pease River group, and the "Blaine of Texas". Each name has advantages and proponents. The Pease River group is adopted here because its upper and lower limits are easily recognized where it crops out on the east side of the Permian Basin. The same cannot be said of the other groups when traced into the area. The Pease River group includes all beds from the base of the San Angelo sandstone up to the base of the Custer which is 10 to 30 feet below the Childress magnesian limestone. The group is disconformable at its base, and a regional unconformity occurs at the top. Average thickness is 870 feet. Surface exposures show many cycles of evaporites, which are now greatly reduced by solution. Except for the area in Nolan County, where red clastics are relatively prominent, the outcrop is uniformly rough with a pro-

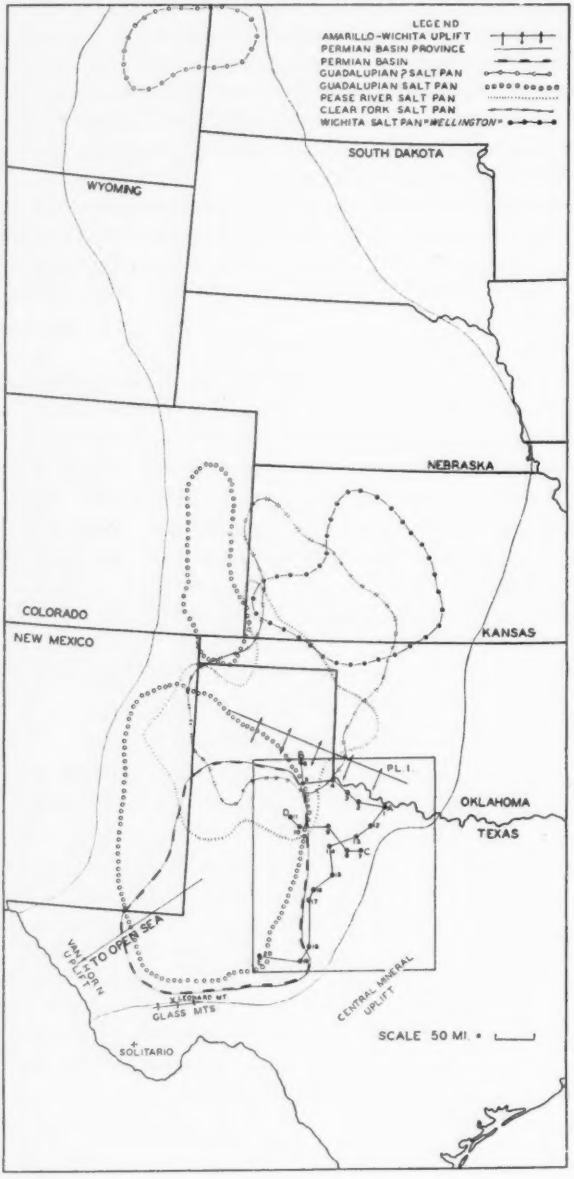

FIGURE 2.—*Permian salt pans*

nounced development of stairlike escarpments, and vegetation is notably different from that of adjacent groups.

In the subsurface clastics gradually diminish basinward, and the recurrence of evaporites is clearly marked, inasmuch as solution has neither removed the sodium chloride nor hydrated the anhydrite.

Figure 2 generalizes the location of the various salt pans in the Permian Basin province. The salt pans are not necessarily massive salt beds but may only represent salty shales. No attempt was made to indicate all the pans found in the Guadalupian.

PALEOGEOGRAPHIC SETTING

The midcontinent as a whole has certain geological characteristics that set it apart from the rest of the United States. Crustal stability has dominated since the beginning of the Cisco. Orogenic movements broke this stability at but one place—the Amarillo-Wichita uplift—which was slightly rejuvenated during post-Permian time. Clastics were derived from the ancient Amarillo-Wichita mountains and form a large percentage of the Wolfcamp and Cisco series. All traces of this positive area were obliterated by burial under the Leonard (Fig. 2).

During Leonard time the Permian Basin province seems to have been a vast peneplain bordered by an upland of slight relief. Except for the widespread San Angelo sandy deposits, clastics coarser than shale are found only on the margins of the Permian Basin. These clastics attain considerable thickness in four main areas: the Anadarko Basin of west-central Oklahoma; Coke and Nolan counties, Texas; Marathon region of West Texas; and the northwestern counties of the Texas Panhandle. The latter area extends westward into New Mexico and probably is related to the Middle Permian Coconino and DeChelly sandstone deposition in Arizona.

Extremely uniform deposition during Leonard time is indicated by lithologic units which range from less than a foot to over 20 feet in thickness and which have been traced for hundreds of miles. Thickening or thinning is of minor importance throughout the area. Seemingly any positive or negative regional movements were epeirogenic, and accordingly small negative movements would cause regional inundation.

All paleontologic and stratigraphic evidence indicates that connection of the Permian Basin with the open sea was toward the south and southwest.

The Llanoria geosyncline, which was filled with thick Ordovician to early Pennsylvanian sediments, extends into Texas from southeastern

Oklahoma. Sediments preserved in this geosyncline have been traced around the east and southeast portion of the Central Mineral uplift of Texas, and thence westward into the Marathon region (Sellards, 1932). Their last appearance is in the Solitario uplift, located in the extreme southwest part of West Texas. A peculiar lithologic facies aids in identification. Black to greenish-black shales associated with coarse and fine sandstones, which in most places are rust brown to greenish gray, overlie a thick succession of cherty limestone. All are now complexly folded and somewhat metamorphosed. Cherty and novaculitic limestone beds, more than 1000 feet thick, dip at high angles and profoundly affect the topography. They provide much of the Permian conglomerates of the Marathon region. The southwesterly open sea connection of the Permian seas must have crossed the sediments of the geosyncline, and because of their geographic position the cherts and novaculites offered effective barriers to an open sea connection, thus affording an ideal area for the growth of bioherms. Permian sediments in this and adjacent areas exhibit a great variety of animal and plant growths whose development would tend greatly to restrict the existing channels to the open sea. The bioherms would also tend to maintain themselves if a negative movement developed.

ENVIRONMENT OF DEPOSITION

The midcontinent apparently was a vast desiccating barred basin during the deposition of the Leonard and, in fact, during most of the Permian. In the southern portion of the Permian Basin there is an area of black shale deposition in which the entire Permian section from Wolfcamp through lower Guadalupian consists of bituminous shales. It is therefore very probable that this area of black shale deposition did not rise above the level of the seas during the time indicated. A few thin brownish-black cherty limestone are scattered through the section. Pyrite is present in many zones. The few sandstones present are very fine, bordering on a grit. Light gray limestones surround the black shale on the east, north, and northwest. The transition from the euxinic environment to one of desiccation is not extremely sharp, for it extends through nearly 60 miles. Continuing in the three directions indicated, dolomites replace the limestones, and farther out these in turn are broken up by interfingering greenish-gray shales and anhydrites. The latter transition is best developed to the north. In the peripheral zone, reddish-brown shales and evaporites dominate the whole section. Thus, outside the "black shale" basin the normal sequence of deposition is gradational, both vertically and horizontally (Pl. 2).

An ideal cycle of chemical deposition of evaporites is as follows:

Sodium chloride (does not crop out because of solution)
Anhydrite (usually hydrated on the outcrop)
Dolomite or magnesian limestone (locally fossiliferous)
Apple-green shale
Blue-green shale
Reddish-brown shale

The thickness of each member of the cycle may vary greatly, but each is usually present. Sand is common in the shale but rare in anhydrite. Cycles are best developed in the Pease River group.

Several hypotheses have been advanced to explain the environment of Permian Basin deposition. One pictures a desiccating sea slowly retreating southwestward. This concept has several drawbacks (Fig. 2). The Leonard seems to have had a wider distribution in the Permian Basin province than any other series except possibly the Guadalupian. Little evidence is found to indicate that the Pease River group was more restricted than either the Clear Fork or Wichita groups. There is a marked tendency for the more soluble evaporites to be confined to an area north of the "black shale" basin. The position of the latter directly in the path of an open sea connection complicates the hypothesis of a southwestwardly retreating sea.

Deposition of black shales of the type found in the Permian Basin requires very specialized conditions that seem to be closely related to the growth of disconnected limestone masses. Mention of bioherms is common in the literature referring to the Permian Basin. The existence of bioherms connotes the presence of lagoons. Lagoonal deposition has been studied from many angles, but explanation is lacking for the many types of deposition found to date. Woolnough (1937) and Twenhofel (1939) have described conditions of deposition remarkably similar to those here included in the Leonard and have advanced the descriptive term barred basin for such deposits.

SECTION A–B

LOCATION OF BORINGS

No. 1.—Humble Oil & Refining Company's Williams & Parker No. 1, Wichita County, Texas. 200 feet from South and East lines of west 213.3 acres, section 25, H. & G. N. R. R. Survey. Elevation, 1029 feet.

No. 2.—Humble Oil & Refining Company's B. D. Barker No. 1, Wilbarger County, Texas. 1650 feet from South and 330 feet from West lines of SE¼ of section 31, block 9, H. & T. C. Survey. Elevation, 1250 feet.

No. 3.—Amerada Petroleum Corporation's J. S. Rice *et al.* No. 1, Hardeman County, Texas. 1635 feet from North and 1679 feet from West lines of section 81, block H., W. & N. W. Survey. Elevation, 1476 feet.

No. 4.—Hines *et al.* W. T. Walling No. 1, Hardeman County, Texas. 1900 feet from the South and 2100 feet from the West lines of section 368, block H., W. & N. W. Survey. Elevation, 1845 feet.

No. 5.—Phillips Petroleum Company's A. C. Hughes No. 1, Hall County, Texas. Center of the N½ of section 4, block H., J. H. Stephens Survey. Elevation, 2225 feet.

No. 6.—H. C. Robinson *et al.* Kuteman No. 1, Donley County, Texas. 660 feet from the North and 330 feet from the West lines of SW¼ of section 88, block 20, H. & G. N. Survey. Elevation, 2704 feet.

DISCUSSION OF SECTION

Section A-B (Pl. 2) parallels the Red River syncline from Wichita County to Hall County, thence north to Donley County (Pl. 1). Boring No. 6 penetrated a rock column more or less characteristic of the succession indicated by borings flanking the eastern portion of the Amarillo-Wichita uplift in the Panhandle of Texas.

The base of the Wichita group in boring No. 1 is not so definite as it is farther south and west. Oil was found at a depth of 685 feet, and this may mark the base of the Wichita group. Where marine strata occur, as in boring No. 2, the base of the Leonard series is readily identified.

Section A-B shows that the base of the redbeds rises westward to Hardeman County, then drops in Hall and Donley counties. It is obvious that the base of the red color does not define a stratigraphic horizon, hence it can be used only locally for structural subsurface mapping. In the northwestern part of the Texas Panhandle, the Leonard has graded almost completely into redbed clastics. The presence of salt in borings Nos. 5 and 6 marks the easternmost recognized occurrence of salt. This salt horizon may be traced into the area from the north and west as it occurs in the Panhandle of Texas, western Oklahoma, and western Kansas. It occurs below the San Angelo horizon and its equivalents and may be correlated with the Stone Corral of Kansas and the Bullwagon of Texas. The Wellington salt of Kansas occupies the lower portion of the Leonard and probably in part is equivalent to the zone of *S. crassitectoria*. All salt beds in the Leonard are essentially free of potash, which distinguishes them from most of the salt deposits in the Custer group and its equivalents.

SECTION C–D

LOCATION OF BORINGS

No. 7.—S. I. Davis' C. C. Lynch No. 1, Throckmorton County, Texas. Center NE¼ of the NE¼ of section 157, B. B. B. & C. Survey. Elevation, 1605 feet.

No. 8.—Amerada Petroleum Corporation's C. J. Kleiner No. 1, Haskell County, Texas. 660 feet from the North and East lines of section 57, J. McCass Survey, A-95. Elevation, 1465 feet.

No. 9.—H. F. Wilcox Oil & Gas Company's W. F. Ross No. 1, King County, Texas. 330 feet from the North and West lines of section 23, R. B. Masterson Survey. Elevation, 1812 feet.

No. 10.—Humble Oil & Refining Company's Matador Land and Cattle Company No. 1, Dickens County, Texas. 1980 feet from the North and East lines of section 26, block 1, Shock and Arnold Survey. Elevation, 2477 feet.

No. 11.—Humble Oil & Refining Company's Matador Land and Cattle Company No. B-1. Motley County, Texas. 330 feet from the East and 7500 feet from the North lines of Blanco County School Lands, A-220. Elevation, 2662 feet.

DISCUSSION OF SECTION

Section C-D (Pl. 2), drawn at right angles to the regional strike (Pl. 1), shows several stratigraphic features of the Leonard. Boring No. 7 is a core test starting just below the Grape Creek beds of the Wichita group. It clearly shows the relation of the anhydrite zone to the base of the Leonard and underlying Elm Creek. This zone does not crop out due to solution by meteoric waters. Rainfall increases toward the east; hence anhydrites are better preserved in the surface expression of the Pease River group (Pl. 1).

The complete absence of redbeds in boring No. 11, except at the San Angelo horizon, is noteworthy. The strata from the base of the San Angelo to the base of the Leonard are almost all dolomite. The cherty dolomites above the San Angelo characterize this portion of the column over much of the Permian Basin but are not known to crop out on the east side. Boring No. 11 encountered thick beds of salt between 1330 and 2030 feet. They belong in the upper part of the Leonard far above the San Angelo horizon and are widespread in the northern portion of the Permian Basin, extending into Oklahoma and Kansas. The overlying Custer group also contains salt beds, but neither the lower nor upper salt beds crop out along the east side of the Permian Basin. The solution of several hundred feet of salt in this region simulates uncon-

formities that are observable in the section above the San Angelo and especially in the several formations constituting the Custer group.

SECTION A–E

LOCATION OF BORINGS

No. 12.—British American Oil Company's J. E. Cope No. 1, Baylor County, Texas. 660 feet from the South and West lines of the NW¼ of section 220, T. & N. O. Survey, A-1369. Elevation, 1318 feet.

No. 13.—H. F. Wilcox Oil & Gas Company's J. E. Fritz No. 1, Baylor County, Texas. 3241 feet from the West and 1573 feet from the South lines of the NE corner of section 193, David Barlow Survey, A-10. Elevation, 1347 feet.

No. 14.—Merry Bros. & Perini, Inc. *et al.* C. L. Williams No. 1, Stonewall County, Texas. Center of SW¼ of SW¼ of section 6, block C., A. B. & M. Survey. Elevation, 1621 feet.

No. 15.—Humble Oil & Refining Company's Dallas Joint Stock Land Bank No. 1, Jones County, Texas. 4458 feet from the South and 990 feet from the West lines of section 267, M. Fragosa Survey, A-176. Elevation, 1804 feet.

No. 16.—Ajax Drilling Company's E. W. Withers No. 1, Nolan County, Texas. 1320 feet from the South and West lines of section 65, block 21, T. P. R. R. Survey. Elevation, 2323 feet.

No. 17.—L. G. Bradstreet's R. V. Kuteman No. 1, Nolan County, Texas. Center of the NW¼ of section 95, block X., T. P. R. R. Survey. Elevation, 2488 feet.

No. 18.—Humble Oil & Refining Company's Lewis & Wardlaw No. 1, Tom Green County, Texas. 2080 feet from the South and 660 feet from the East lines of section 5, block 17, H. & T. C. Survey. Elevation, 2142 feet.

No. 19.—Stanolind's Williams No. 1, Irion County, Texas. 1320 feet from the East and South lines of section 1196, T. & T. R. R. Survey. Elevation, 2544 feet.

No. 20.—Big Lake's University No. 17-C, Reagan County, Texas. 2528 feet from the South and 2556 feet from the West lines of section 25, block 9, University Lands Survey. Elevation, 2678 feet.

DISCUSSION OF SECTION

Section A-E (Pl. 2) parallels the strike of Permian outcrops from Wichita County to Tom Green County (Pl. 1). Included is a composite

type section of the outcrop on which are indicated the more commonly recognized geologic names. The section continues westward from Tom Green County to Reagan County.

Although fossils are lacking to the north and east yet correlations of the stratigraphic units in boring No. 18, Tom Green County, are relatively certain. So many wells have been drilled to the north and east that a well-to-well lithologic comparison brings out all essential details. Boring No. 18 starts on Comanchean beds. The top of the zone of *Perrinites* is at 250 feet. Regional mapping indicates that no Custer sediments intervene between the Comanchean and the zone of *Perrinites*. The base of the San Angelo was encountered at 605 feet. First fusulinid fragments were noted at 1200 feet. The zone of *S. crassitectoria* starts at 1800 feet, and the base of the zone of *Perrinites* is at 2410 feet. First fusulinids of the zone of *Properrinites* were noted at 2720 feet and are very abundant between 3000 and 3100 feet.

Boring No. 19 is not easy to correlate by lithology; accordingly paleontologic evidence must be sought. No single boring generally yields as much information concerning fossils as is necessary; hence studies on adjacent borings were used in making these correlations. Boring No. 19 starts on the Comanchean and at 650 feet enters the zone of *Waagenoceras*. No sediments of Custer age intervene between the base of the Comanchean and the zone of *Waagenoceras*. At 1230 feet *Parafusulina rothi* and *Parafusulina sellardsi* appear. These fusulinids are diagnostic of the lower and middle Delaware (Fig. 1). *Parafusulina* sp., which seems to be more highly developed than fusulinids in the underlying zone of *Perrinites*, has been found at 3030 feet. The top of the zone of *Perrinites* is placed at 3320 feet. The San Angelo probably is equivalent to the sandy zone between 3800 to 3850 feet. Fusulinids characterizing the zone of *Perrinites* occur between 4170 to 4200 feet. The top of the zone of *S. crassitectoria* is at 5300 feet, and the first fossils indicating the zone of *Properrinites* occur at 5920 feet.

Boring No. 20, in Reagan County, has the following section: Comanchean from the surface to 510 feet; Dockum to 730 feet; Elk City sandstone to 880 feet; base of Doxey shale 930 feet; base of the Whitehorse of Texas at 2930 feet. The base of the recently proposed Ochoa series is at a depth of about 1500 feet. In this boring sediments of Custer age rest upon the middle portion of the zone of *Waagenoceras*. Cores of strata between 3100 and 3200 feet are composed of *Parafusulina lineata*, *P. maleyi* var. *referta* and *P. rothi*, indicating a fusulinid bioherm. These fossils suggest that the strata between 2930 and 3900 feet are equivalent in age to the zone between 650 and 1800 feet in

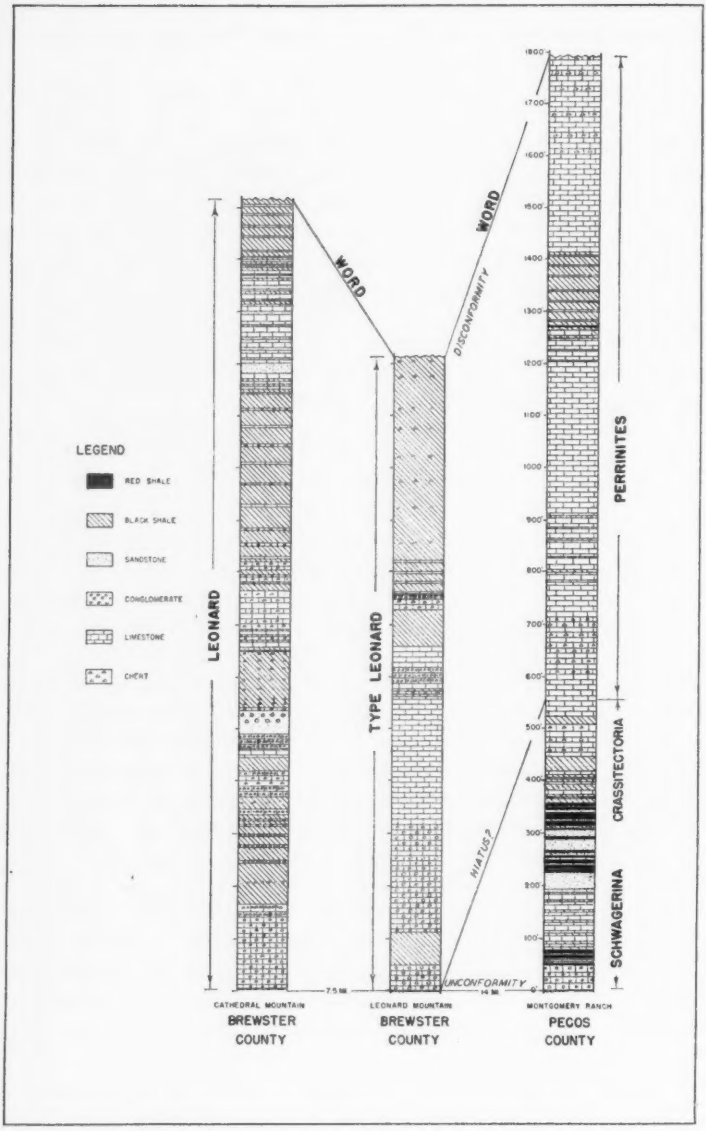

FIGURE 3. —*Cross section, type Leonard of Glass Mountains, West Texas*

boring No. 19. In boring No. 20, the zone of gritty gray-black shales gives way to black shales and cherty limestones at 4810 feet, which is considered the top of the zone of *Perrinites*. This contact is much the same as on the outcrop in the Glass Mountains. Determination of the base of the zone of *Perrinites* is problematical, but it is placed at 6290 feet, at which depth commercial oil production has been found in several wells in the Big Lake field. Inasmuch as oil has been found at or near unconformities, it is probable that this producing horizon marks the contact between the zones of *Perrinites* and *Properrinites*. Fusulinids found just below 6550 feet are too fragmentary to identify. If the above contact is correctly placed, the thickness of the zone of *Properrinites* is about 1700 feet. A number of borings, for which paleontologic evidence is available, have penetrated an equal or greater thickness of Wolfcamp beds. Since boring No. 20 is in the heart of the "black shale" basin, no subdivisions of the zone of *Perrinites* are attempted.

TYPE LEONARD OF GLASS MOUNTAINS, WEST TEXAS

Three measured sections (Fig. 3) of the outcrop at the type area of the Leonard in the Glass Mountains show the lithology of the *Perrinites* zone. Redbeds can be traced for several miles along the northeast part of the Glass Mountains scarp. They are wholly within the zone of *S. crassitectoria*. The indicated thickness of the zone of *Perrinites* in the type area is about half as much as it is along the east side of the Permian Basin.

The comparatively uninhabited nature of the Marathon region makes the location of geographic landmarks difficult. The general locations of the three measured sections used in the cross section (Fig. 3) are as follows: Montgomery Ranch section, Pecos County, is measured north-northwest from Montgomery Ranch in Hess Canyon Quadrangle, Marathon region. This section is 14 miles northeast of the Leonard Mountain section. The Type Leonard section, Brewster County, is measured north-northwest from the top of Leonard Mountain to Road Canyon, thence north to the summit of the north-dipping cuesta. It is 7½ miles northeast of Cathedral Mountain section, Hess Canyon quadrangle, Marathon region. Cathedral Mountain section, Brewster County, starts on the bolson flats of the Marathon plain and is measured N. 15° W. to the top of Sullivan Peak in the Cathedral Mountains, Altuda Quadrangle, Marathon region.

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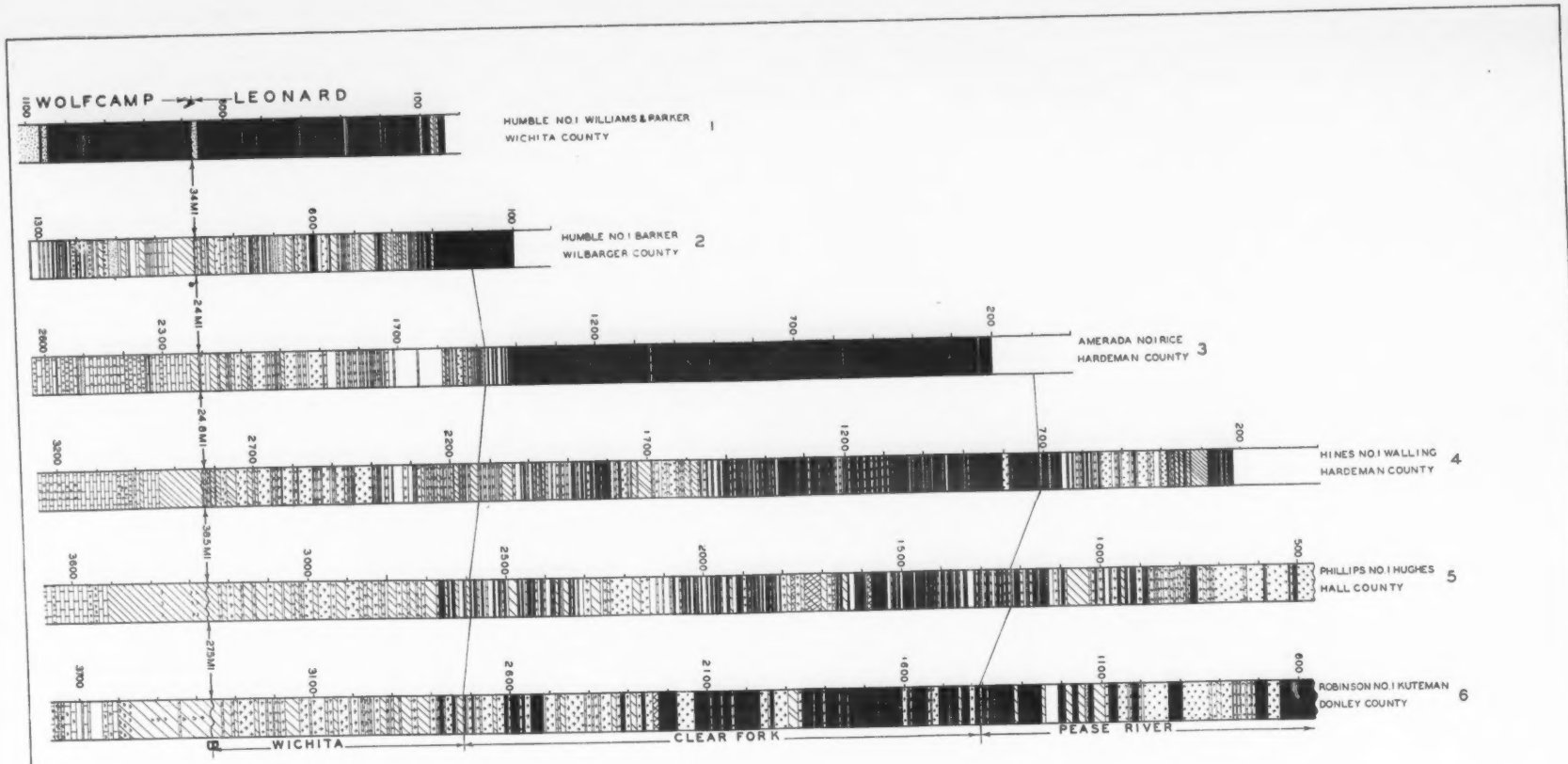
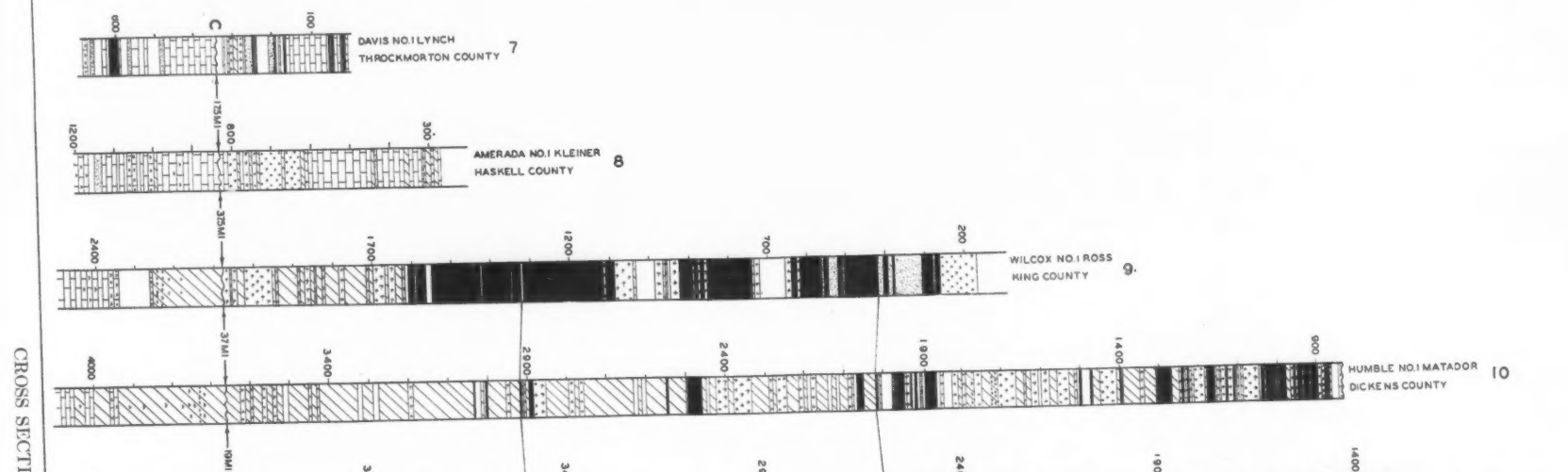
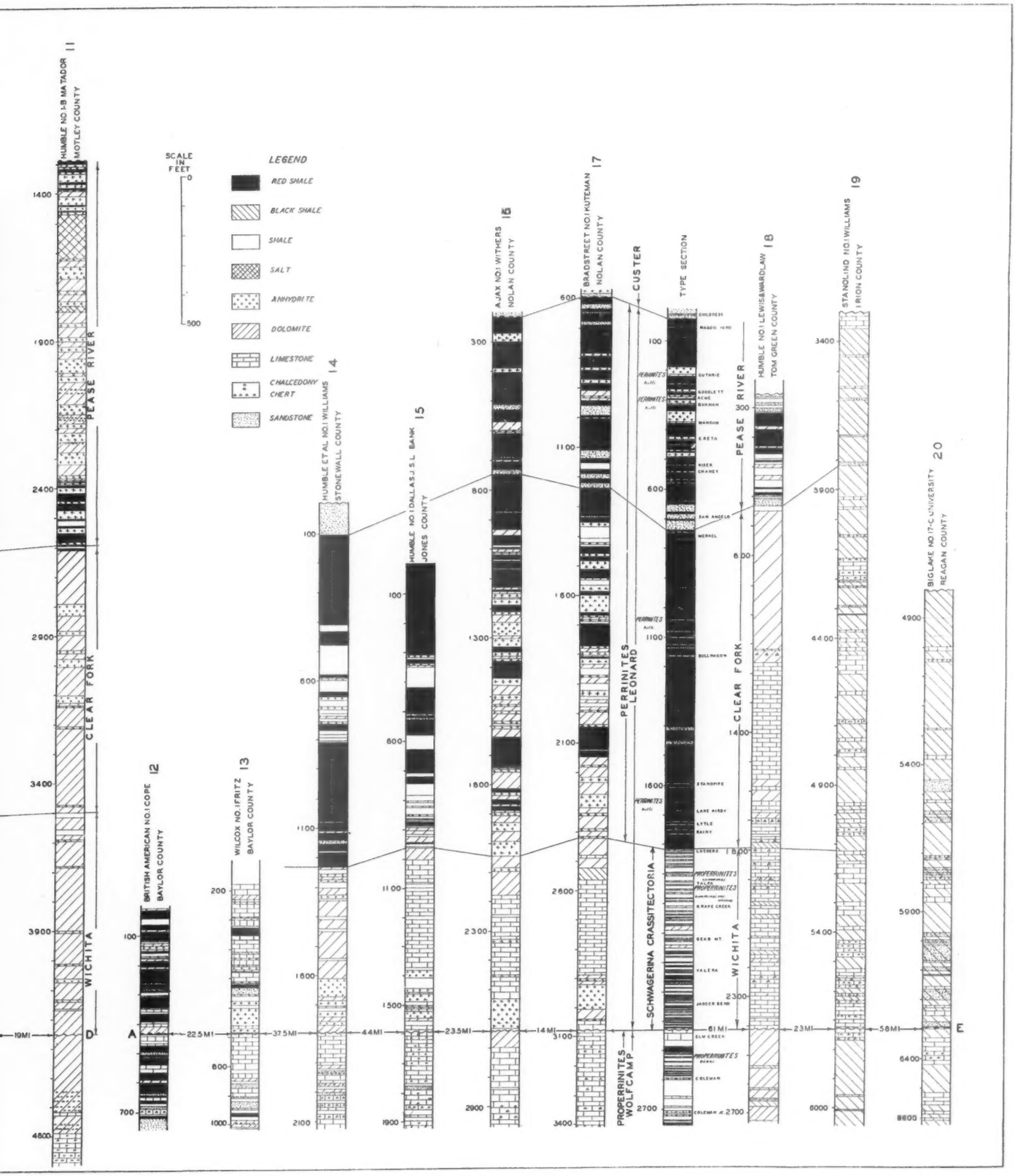




CROSS SECTIONS



ECTIONS A-B, C-D, AND A-E